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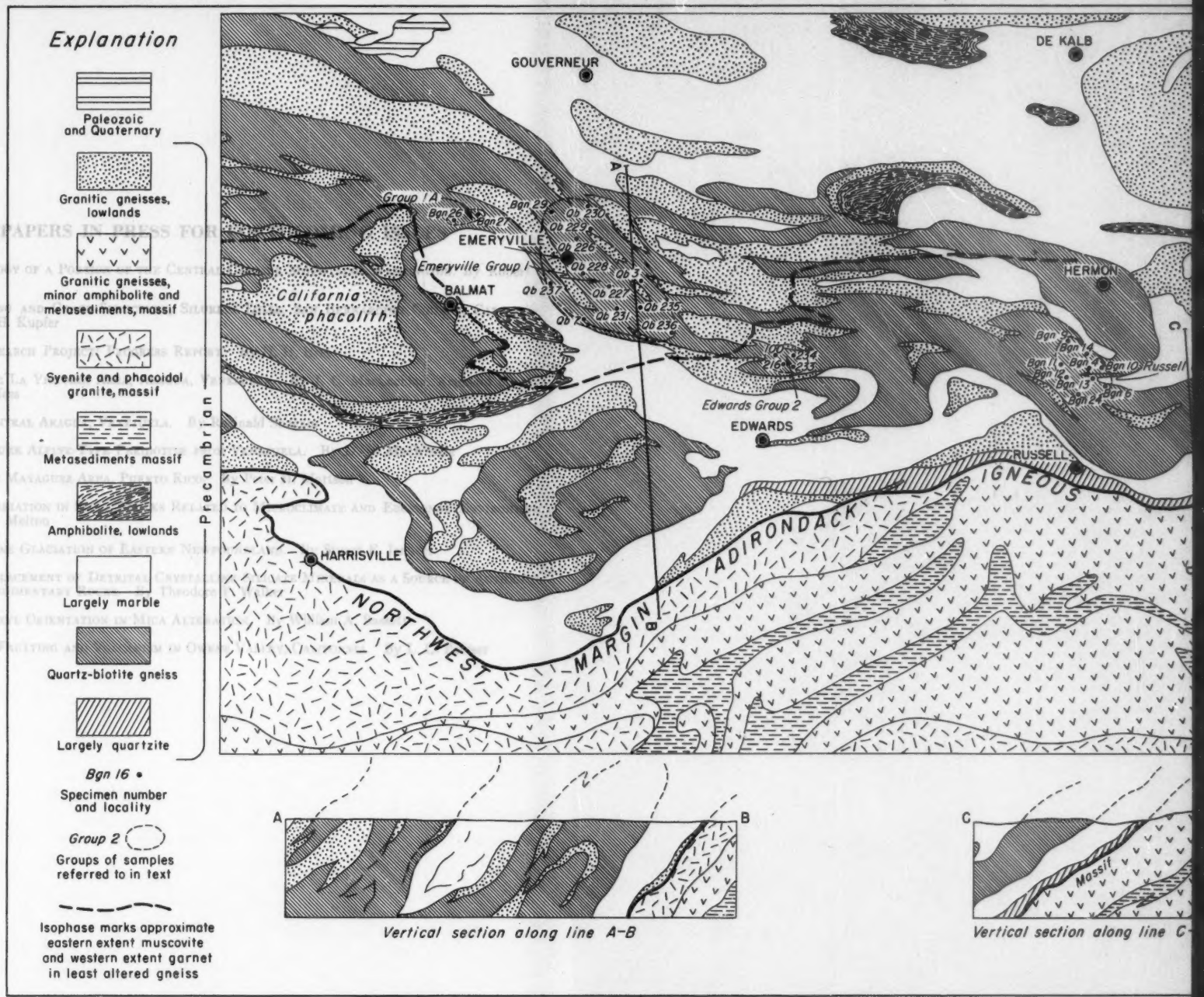

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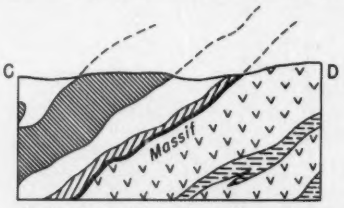

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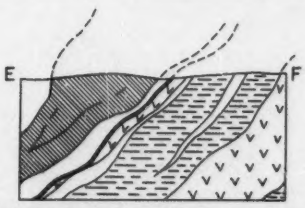

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 71, PP. 1-58, 12 FIGS., 3 PLS. JANUARY 1960

PROGRESSIVE METAMORPHISM AND GRANITIZATION OF THE MAJOR PARAGNEISS, NORTHWEST ADIRONDACK MOUNTAINS, NEW YORK

PART II: MINERALOGY

BY A. E. J. ENGEL AND CELESTE G. ENGEL

(WITH ANALYSES OF TRACE ELEMENTS BY
A. A. CHODOS AND ELISABETH GODIJN)

ABSTRACT

Changes in the kinds and properties of minerals in the Adirondack paragneiss are related to its progressive metamorphism. The least altered gneiss at Emeryville, New York, consists of quartz, biotite, and oligoclase, with small amounts of muscovite and microcline. This assemblage formed at about 500° C under a lithostatic load of about 5 miles. As the least altered geneiss is traced northeastward into what was a deeper, hotter environment of metamorphism the biotite and muscovite react to produce garnet; K_2O, SiO_2, and water are expelled. At Colton, New York, 35 miles northeast of Emeryville, temperatures of metamorphism probably reached 600° C. There, the least altered gneiss consists of quartz, calcic oligoclase, biotite, garnet, and accessory K feldspar. Migmatitic and granitized paragneiss throughout the region is enriched in K feldspar and depleted in biotite, garnet, and plagioclase.

The microcline in least altered gneiss at Emeryville occupies 1–5 per cent of the rock. It is distinctly perthitic and contains up to 2 weight per cent Na_2O, 0.2 to 0.3 per cent BaO, and from 300 to 2600 parts per million Sr. With increasing grade of metamorphism gridiron twinning disappears (orthoclase appears?), and Na_2O content increases to 2.4 per cent.

The K feldspar in the migmatite and granitized gneiss throughout the Emeryville–Colton region is almost wholly a perthitic microcline, with some abrupt and local variations in amounts of Na, Ba, and Sr. These variations in alkalis occur within individual bodies of granitic gneiss, and between microcline in these bodies and the microcline porphyroblasts in the enveloping paragneiss.

Plagioclase in the least altered gneiss at Emeryville averages Ab 73, forms about 40 per cent by volume of the rock and shows a separation of $2\theta(1\bar{3}1)$ and $2\theta(131)$ of 1.61. Only 5 per cent of the plagioclase is visibly twinned (albite and pericline types). Most grains are sercitized. Zoning is very rare. With increasing grade of metamorphism plagioclase in the least altered gneiss becomes more abundant, coarser-grained, and profusely twinned (largely albite twinning). At Colton, plagioclase forms half the rock and is Ab 67 with $2\theta(1\bar{3}1) - 2\theta(131)$ of 1.71.

Plagioclase in migmatite and granitized gneiss decreases in volume and Ab content as K feldspar content of these rocks increases. The average granitic gneiss at Emeryville contains about 30 per cent plagioclase of Ab 86.

A greenish-brown, Al-rich biotite constitutes 18 per cent of the least altered gneiss at Emeryville. With progressive metamorphism biotite decreases in volume, in size of unit cell, and in its content of Fe_2O_3, FeO, OH, Sc, and Mn. Grain size increases, as do amounts of TiO_2, MgO, Ba, Cr, F, V, and probably in Ca. Color changes to deep reddish brown as a function of the increasing ratio of TiO_2:FeO. N_x and density undergo small, less consistent changes. The 1M polymorphs predominate, but 2M forms are not uncommon. There is no correlation of polymorphic type with grade of metamorphism. Granitization of gneiss results in the decomposition of biotite, which forms less than 10 per cent by volume of the average granitic gneiss. Biotite of the incipiently granitized gneiss is very similar in composition to that in associated least altered gneiss.

At Emeryville garnet is confined to pegmatites and migmatitic gneiss, but from Edwards northeastward it occurs throughout the least altered gneiss as a product of regional metamorphism. The garnets associated with pegmatite near Emeryville are spessartitic almandites with 6 per cent MnO, 4 per cent MgO, and 2 per cent CaO. The

first garnets to appear throughout least altered gneiss (at Edwards) contain less than 2 per cent MnO and about 6 per cent MgO. As the grade of metamorphism increases the garnets increase in abundance and in per cent MgO, Cr, Ti, V, and Zn. Complementary decreases occur in dimensions of the unit cell, in index of refraction, density, Fe_2O_3, FeO, MnO, Ba, Sc, Y, and Yb.

In the selective distribution of mafic elements in the paragneiss, biotite accumulated more Mg, Ba, Co, Cr, Cu, Ni, Ti, and V than did garnet. Garnet is enriched in Mn, Fe, Ca, Sc, Y, and Yb. With increasing grade of metamorphism the ratio Fe^{++}+ Mn:Mg in garnet decreases from 8 to 5, and in biotite from 2.8 to 1.7 Fe^{+++}:Fe^{++} decreases in both biotite and garnet.

SiO_2:Al_2O_3 decreases in both total rock and in plagioclase with increasing grade of metamorphism of least altered gneiss. K_2O is concentrated in biotites and K feldspar, and Na in plagioclase. With progressive metamorphism K_2O:Na_2O decreases in K feldspar and total rock.

The pronounced and systematic variations in concentrations and ratios of elements in minerals and total rock between Emeryville and Colton are marked by only one classical isograd. This is the reaction of quartz, biotite, magnetite, and muscovite to produce garnet. For this region, and perhaps many others, the scope and patterns of progressive metamorphism are shown in much greater detail by changes of ratios of elements and oxides in minerals. The ratio of two easily determined oxides that have opposing trends in concentration is especially definitive. TiO_2:MnO in biotites changes from 16 to 520 between Emeryville and Colton. TiO_2:MnO in garnets changes from .003 at Emeryville to .06 at Colton. Ratios of Fe:Mg, Mn:Mg, or Fe^{++}+ Mn:Mg in both minerals are even more informative but also more difficult to obtain accurately.

The mineral assemblages in the paragneiss at Emeryville are complemented by diopside in adjoining dolomites, sillimanite in nearby pelitic schists, and hornblende-andesine in amphibolite interlayers. At Colton, diopside persists in the dolomitic marble, whereas in the amphibolite ortho- and clinopyroxenes have formed at the expense of about half the hornblende. These relationships indicate the transition from upper amphibolite to lower granulite facies and are accompanied by marked dehydration, decarbonation, and a basification of the total rock.

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Introduction

This report discusses the mineralogy of the Adirondack paragneiss, with special emphasis on the interrelationships between mineral properties and those of the total rock, its geologic setting, and grade of metamorphism. A discussion of the "total rock" has preceded this as Part I of these studies and was published in 1958. These two papers are in turn complementary to an earlier discussion of the regional relationships of the Adirondack paragneiss, its general geological features, and its possible origin (Engel and Engel, 1953).

The Adirondack paragneiss consists of quartz, biotite, plagioclase, and garnet with locally some muscovite and K feldspar; accessories are zircon, apatite, sphene, magnetite, ilmenite, and pyrite. Its bulk composition is about that of an average graywacke (Pettijohn, 1957, p. 307; Engel and Engel, 1953; 1958a). Biotite, plagioclase, garnet, and K feldspar have been analyzed, X-rayed, and studied optically. The properties of each mineral, especially chemical composition and the partitioning of elements among the minerals, show marked systematic variations with changes in grade of metamorphism. These variations offer definitive indices of the metamorphic facies and amplify concepts of petrogenesis.

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Geologic Setting

The paragneiss is part of a thick (12,000-16,000 feet) metasedimentary sequence of Grenville-like rocks in the northwest Adirondacks (Engel and Engel, 1953). This sequence extends northeastward obliquely across the Grenville Lowlands for about 75 miles. The northeasterly 50 miles of the belt of paragneiss is shown on Plate 1. Within this region the paragneiss appears to be grossly monoclinal, but overturned to form the southeast flank of a regional anticlinorium. Two siliceous marble zones lie northwest of and stratigraphically below the gneiss. A third siliceous marble lies to the southeast, stratigraphically above it. Thin amphibolite interlayers are common within the gneiss. The northeast-trending gneiss, amphibolites, and marble converge at low angles with the trace of the arcuate perimeter of the Central Adirondack Massif in the vicinity of Colton, New York (Pl. 1). There, temperatures of metamorphism in the metasedimentary sequence are thought to have reached 600°–625° C. Southwest along the strike, away from Colton, maximum temperatures of metamorphism appear to have decreased uniformly to the Emeryville area, where maximum tem-

PLATE 2.—PHOTOMICROGRAPHS OF LEAST ALTERED GNEISS AT EMERYVILLE, NEW YORK

FIGURE 1. Quartz-biotite-oligoclase gneiss (Qb 7) under uncrossed nicols (× 200)
FIGURE 2. Quartz-biotite-oligoclase gneiss (Qb 7) under crossed nicols (× 200)

FIGURE 1

FIGURE 2

PHOTOMICROGRAPHS OF LEAST ALTERED GNEISS AT EMERYVILLE, NEW YORK

FIGURE 1

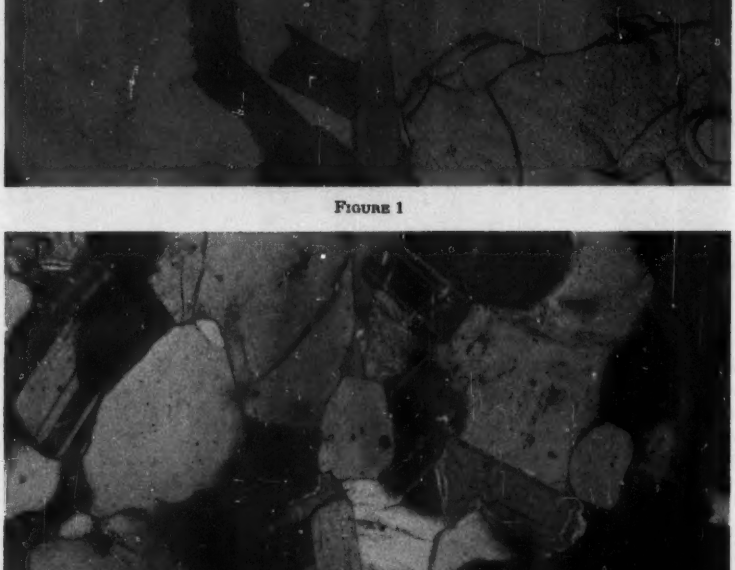

FIGURE 2

PHOTOMICROGRAPHS OF LEAST ALTERED GNEISS AT COLTON, NEW YORK

peratures of about 500° C are recorded (Part I, Engel and Engel, 1958a). Beyond Emeryville to the southwest and west, maximum temperatures of metamorphism appear to have been essentially the same, about 500° C, although, at least locally, small gradients undoubtedly existed.

Host Rock

The paragneiss is a complex of more and less granitized and migmatized rock. As explained in the discussion of the "total rock" (Engel and Engel, 1958a), the least altered gneiss in the Emeryville area—at the lower-temperature end of the belt of progressive metamorphism—is thought to represent the approximate chemical composition of the parent sediment (Fig. 1). In any given section or area the least altered parts of the gneiss are remarkably monotonous in texture and in mineralogical and chemical composition. Mineralogically, the least altered gneiss at Emeryville is a quartz-biotite-oligoclase-muscovite gneiss averaging quartz 38, plagioclase (Ab 73) 40, K feldspar 3, biotite 18, muscovite 1 (Pl. 2). As the least altered gneiss is traced northeast (with increasing temperature of metamorphism) toward Colton the mineral composition changes as follows: muscovite disappears, garnet appears, plagioclase increases in abundance and in average An content, and biotite and quartz decrease (Fig. 1). Grain size increases, and the rock is "basified" by depletion of K, Si, Fe^{+++}, H_2O, and Ba, and by a progressive increase in Al, Fe^{++}, total Fe, Mg, Ca, Cr, Ga, Ni, and V.

This basification of the gneiss appears to be a metamorphic (ultrametamorphic) process that is obvious at about 550° C (Russell, New York, Pl. 1) and strongly defined at 600° C (Colton, Part I, Engel and Engel, 1958a, Fig. 2). The mobilized Si, K, and H_2O appear to have been partly liberated, and partly trapped as venitic migmatite.

Granitization and migmatization[1] of parts of the gneiss are accompanied by an increase in K feldspar and Ab content of plagioclase and by a decrease in biotite, plagioclase, and quartz. Chemical changes in major elements include an increase in K and a decrease in Ti, Fe, Mg, Ca, H_2O, and in Na:K ratio. All granitizing substances in the gneiss in areas of lower-temperature metamorphism appear to be introduced either laterally or from below. Those in areas of highest-temperature metamorphism are partly introduced, partly derived locally from the gneiss (Part I, Engel and Engel, 1958a).

Metamorphism of the gneiss complex to its intermediate and high rank occurred approximately 1 billion years ago, during the emplacement of the bulk of Adirondack granitic rocks. The age of the parent sediment is unknown. Its post-metamorphic thickness is about 2000 feet.

Choice and Location of Samples

Each of the minerals discussed is from the specimens of paragneiss whose compositional properties are discussed in Part I. Locations of the samples are shown in Plate 1 and represent the most uniform distribution permitted by field relations, texture, and retrograde alteration. These problems are discussed in Part I, pages 1376–1377.

Presentation of Mineral Data

A standard method of presentation is used to aid comparisons of the analytical data on various figures and tables. Inherent in the method is the conclusion that a positive temperature gradient existed in the gneiss between Emeryville and Colton during metamorphism. The relative temperature of formation of any specimen in this study may be indicated by the distance of the sample locality from the edge of the massif, as measured along a line on the earth's surface drawn normal to the trace of the massif. This measurement is used as a scale (horizontal co-ordinate) and calibrated against an approximation of the temperature of crystal growth (reconstitution) in Figure 2 of Part I. This scale seems justified by the fact that the outer surface of the massif and the southeast, overturned side of the gneiss are subparallel and dip at moderate angles northwestward (Pl. 1). In those figures indicating properties of minerals as a function of progressive metamorphism the horizontal co-ordinate is used

[1] The words granitization and migmatization are used in a purely descriptive sense. (*See* Engel and Engel, 1958a, p. 1371.)

Plate 3.—PHOTOMICROGRAPHS OF LEAST ALTERED GNEISS AT COLTON, NEW YORK

Figure 1. Quartz-biotite-oligoclase-garnet gneiss (Bgn 18) under uncrossed nicols.

Figure 2. Quartz-biotite-oligoclase-garnet gneiss (Bgn 18) under crossed nicols.

as a measure of the shortest horizontal distance of the center of each group of specimens from the massif. The vertical co-ordinate represents the property of the mineral (or assemblage of minerals) being compared.

The specimens of Group 1 (Emeryville area) are rather uniformly distributed around a point 9.7 miles from the perimeter of the massif, as measured along a line on the map drawn essentially normal to the perimeter (Pl. 1). Specimens of Groups 2, 3, 4, and 5 are distributed around points respectively 4.7, 3.7, 2.7, and 1.5 miles from the perimeter of the massif (Pl. 1). Comparisons are confined for the most part to statistical values of each group rather than to those of individual specimens. Consequently, group values for given properties of rocks or minerals in the Emeryville area (Group 1) are plotted at a point on the horizontal co-ordinate 9.7 miles from the right-hand side of the graph (perimeter of the massif). Group values for rocks or minerals in the Edwards area (Group 2) are plotted at a point scaled 4.7 miles from the right side (massif), and so on. Considering the complexity of the geology, comparisons of properties of individual rocks or component minerals within any single group as a function of slight differences in distance from the massif do not seem justifiable. Hence, the relative position of a specimen within the group is not to be regarded as highly significant, although for simplicity the sequence of specimen numbers is kept constant throughout the illustrations.

Sample numbers of constituent minerals are the same as for the rock. There are five groups of minerals from least altered and granitized gneiss, and they are designated as follows:

Emeryville, Group 1. Samples in the Emeryville area, at the southwest end of the 35-mile segment of gneiss. The center of the group is approximately 9.7 miles from the nearest margin of the massif (Pl. 1). The group consists of 12 rock specimens. Four of these specimens—Qb 7, Qb 230, Qb 231, and Qb 228— are from the cores of the larger areas of least altered gneiss. The remaining 8 specimens are examples of gneiss in various stages of granitization and *lit-par-lit* injection.

Group 1A (West Balmat area) consists of 2 rock specimens from the gneiss complex. They were taken just southwest of Group 1 between it and the California phacolith (Pl. 1). Specimen Bgn 27 is from the core of a large lens of least altered gneiss, whereas Bgn 26 is taken from a zone of subtle transition between this gneiss and equigranular, uniformly feldspathized gneiss. The close proximity of specimens of Group 1 and Group 1A (Pl. 1) is reflected in the similar features of texture and mineral composition. In most respects these two groups may be considered as one, and the total number of groups as 5.

Group 2 (Edwards area) consists of 4 specimens. The center of Group 2 is approximately 4.7 miles from the nearest perimeter of the massif and about 6.7 miles northeast of the center of Group 1; their opposed edges are about 3 miles apart (Pl. 1). Specimen Qb 234 of Group 2 is from the center of the largest body of least altered gneiss in the Edwards area. Specimen Qb 100 is an inequigranular augen gneiss, and Qb 216 and Qb 233 are equigranular, uniformly feldspathized gneiss.

Group 3 (Russell area) consists of 9 rock specimens; 4 of these were studied in detail. The center of Group 3 is about 3.7 miles from the massif and about 9 miles northeast of the center of Group 2. The opposed perimeters of these groups are about 8 miles apart as measured along the strike. Specimen Bgn 14 is from the center of a lens of least altered gneiss. The other three rock specimens studied in detail are from a zone of transition between this gneiss and a faintly foliated, inequigranular granite gneiss.

Group 4 (Pierrepont area) consists of 2 specimens of gneiss studied in detail. This group has its center about 11 miles northeast of the center of Group 3 and about 2.7 miles from the perimeter of the massif. Opposed boundaries of these groups are 10 miles apart. No specimens of least altered gneiss suitable for mineral separations were obtained in the Pierrepont area. Specimen Bgn 25 represents incipiently feldspathized gneiss from a zone that merges into equigranular granitic gneiss. Specimen Bgn 16 represents rock intermediate between the incipiently feldspathized gneiss and the inequigranular granite gneiss.

Group 5 (Colton area) consists of 4 specimens of least altered gneiss—Bgn 18, Bgn 19, Bgn 20, and Bgn 21—and 2 specimens of uniformly granitized gneiss, Bgn 22 and Bgn 23. The center of this group is about 1½ miles from the Adirondack massif, and 3 specimens are within a mile of the massif (Pl. 1). The center of Group 5 is about 7 miles northeast of the center of Group 4. Limitations of time prevented a thorough chemical study of more granitized gneiss from this area although several analyses of such specimens were made (Part I, Engel and Engel, 1958a, Tables 12, 13).

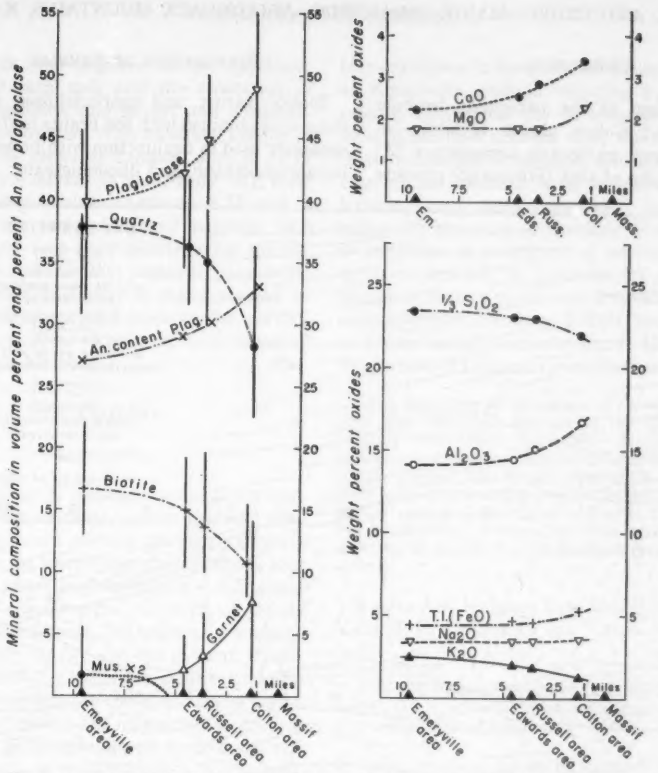

FIGURE 1.—GENERALIZED CURVES INDICATING CHANGES IN MINERAL AND MAJOR-ELEMENT COMPOSITION OF LEAST-ALTERED GNEISS AS TRACED FROM EMERYVILLE TO COLTON
Curves are based on data from 73 specimens.

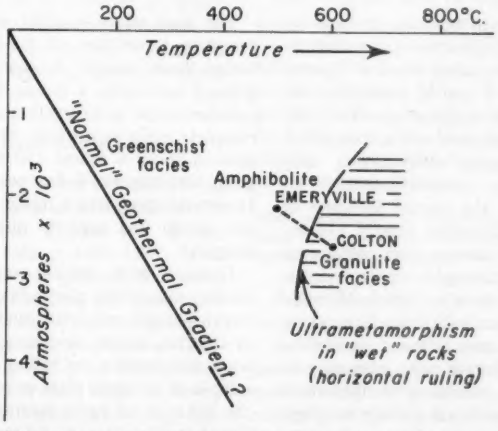

FIGURE 2.—MINERAL ASSEMBLAGES IN THE EMERYVILLE-COLTON REGION PLOTTED IN RELATION TO THE PRINCIPAL METAMORPHIC FACIES, TO *T*, *P*, AND ULTRAMETAMORPHISM
Fluid pressures not appreciably less than load pressures in this phase of Adirondack metamorphism.

Assumptions

Metamorphism of the paragneiss between Emeryville and Colton almost certainly involved a positive gradient in temperature (T) and probably one of load (lithostatic) pressure

Preparation of Samples

Biotite, garnet, and quartz-feldspar separations were obtained with the Frantz isodynamic separator used in conjunction with bromoform, tetrabromoethane, and diiodomethane.

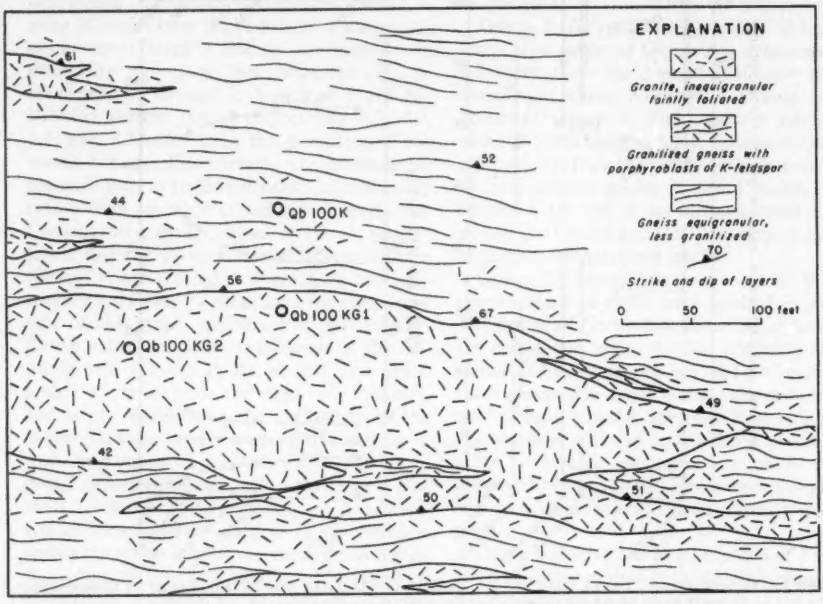

FIGURE 3.—FIELD RELATIONSHIPS OF GRANITE, AUGEN GNEISS, AND EQUIGRANULAR PARAGNEISS NEAR EDWARDS, NEW YORK, AND THE LOCATIONS OF SPECIMENS Qb 100K, Qb 100KG1 and Qb 100KG2

(P_l). The load pressure presumably was to some degree nonhomogeneous—i.e. stress is involved, possibly decreasing toward Colton, the area of maximum T and of widespread ultrametamorphism and magmation. The rise in T (and P_l?) northeastward was accompanied by (caused?) increasing dehydration and decarbonation of the metasedimentary sequence, and probably the partial pressures of water and CO_2 (P_f) increased toward Colton.

The composite of known and ambiguous parameters in the metamorphic environment is frequently referred to here as "grade of metamorphism," which increased from Emeryville to Colton. The abscissae of most succeeding diagrams, read from left to right, also may be thought of as this increasing metamorphic grade. Figure 2 represents an attempt to graph the relations envisaged in the conventional context of regional metamorphism involving T, P, and ultrametamorphism.

The most effective grain size for essentially complete separation of constituent minerals varied from sample to sample. In coarser-grained specimens a minus 100 and plus 150 (meshes to the inch) fraction achieved virtually complete grain separation, whereas in the fine-grained rocks a minus 150 and plus 200 size group was required to free component minerals. In several specimens a minus 200 and plus 300 size group was used to obtain the purified sample.

Garnets were readily separated, whereas biotite, muscovite, plagioclase, and K feldspar were separable only with great difficulty. Traces of chlorite, zircon, and apatite were removed from the biotites by rolling the nearly clean sample on an agate plate or by gently pounding the flakes in an agate mortar. The biotite was jigged on glazed paper and rerun through heavy liquids. A similar procedure was employed in purification of muscovite. Commonly some 70

to 80 hours was required in the combined crushing of each rock and the separation of biotite, muscovite, and garnet and a quartz-feldspar rich fraction. The resultant sample of biotite rarely exceeded 6 grams, and, in 12 of the biotites analyzed in this study, the total "clean" sample ranged between 1.75 and 4 grams. These samples ranged between 97.0 and 99.5 per cent pure biotite, with chlorite and plagioclase as the principal impurities. The relatively small size of many samples of biotite required a modification of the conventional wet chemical techniques, as is discussed in Analytical Methods.

K Feldspar

Introduction

Studies of the K feldspars in the gneiss complex have been confined to their chemical composition and more obvious features viewed in thin section. At least 80 per cent of the K feldspar in the entire gneiss complex is microcline, more than three-fourths of which is clearly perthitic, with abrupt and seemingly irregular variations in Na, Sr, and Ba content. Essentially all the K feldspar in migmatite and pegmatite is microcline. In the least altered gneiss, however, the clarity and abundance of gridiron twinning in K feldspar appear to decrease with increased rank of metamorphism of the rock. This decrease may indicate a decrease in "microcline domains"—the existence of "orthoclase"?—at Colton. The often-stated conclusion that porphyroblasts of K feldspar in augen gneiss adjacent to granite are "identical" in composition and other properties to those in the granite does not apply in the Adirondack paragneiss.

K Feldspar in Least Altered Gneiss

K feldspar commonly forms less than 5 per cent of the volume of least altered gneiss in the Emeryville-Colton area (Engel and Engel, 1953, p. 1061–1064; 1958a, p. 1393–1403). Most of this K feldspar is fine-grained (0.2 mm), fresh, very irregular in outline, and appears as "fillings" in a framework of coarser grains of plagioclase, biotite, garnet, and quartz. Gridiron twinning is faintly visible on perhaps half the grains; the remainder show slight to clearly defined undulatory extinction, which may represent occult twinning.

In general, there appears to be a slight predominance of K feldspar having only undulatory extinction in the gneiss at Colton, whereas at Emeryville gridiron twinning is well developed throughout the K feldspar. Possibly these differences coincide with the occurrence of orthoclaselike K feldspar, at Colton. Relations similar to these, in K feldspars of other terranes, have been interpreted to indicate either (1) inversion (transition) of microcline to orthoclase as a function of increasing grade of metamorphism, or inversely (2) an alteration of orthoclase to microcline as the grade of metamorphism decreases. Eskola, for example, in discussing differences in the K feldspars in the gneisses of Lapland (granulite facies), notes:

> "The first stage of alteration of the clear orthoclase into microcline appears in the form of undulating extinction. Gradually extremely fine twinning lamellae appear in the most strained parts of the crystals and finally the whole crystal is cross-hatched microcline, while other specimens display only well developed quadrille structure—The development of microcline from strain shadows seems to be specific for the granulite facies" (1952, p. 148).

A suite of K feldspars from Lapland has been X-rayed by Goldsmith and Laves who comment:

> ". . . microclinization described by Eskola can be shown by X-ray diffraction to go hand-in-hand with variability of triclinicity" (1954b, p. 110).

In some instances the entire range from monoclinic orthoclase to completely triclinic microcline seems to exist in single hand specimens of gneiss (Goldsmith and Laves, 1954b, p. 110).

Following the work of Goldsmith and Laves, including heat treatments of K feldspars (Goldsmith and Laves, 1954a), Heier X-rayed metamorphic K feldspars from southern and northern Norway. Heier, following Goldsmith and Laves, used the difference in the spacing of the (131) and ($1\bar{3}1$) lines as a measure of triclinicity (microclinization). Heier (1957, p. 468) concludes that:

> "the orthoclase $\rightleftharpoons$ microcline transition takes place very close to the conventional granulite facies—amphibolite facies transition".

Heier infers that the temperature of this transition is about 500° C.

As Goldsmith and Laves (1954a) have suggested, however, the degree of triclinicity or monoclinicity (microcline $\rightleftharpoons$ orthoclase) of naturally occurring K feldspars may be profoundly influenced by the thermal and deformational histories peculiar to a given terrane. It seems quite possible that variations may exist

between the grade of metamorphism (as indicated by either mineral assemblages or maximum temperature of metamorphism) and the structural state of the K feldspar.

Additional studies of the structure and ordering of the K feldspars in the least altered gneiss would seem a desirable project although problems of concentration of grains from homogeneous specimens of the least altered gneiss are most imposing.

In the garnetiferous segments of least altered gneiss—Edwards to Colton—the scattered grains of K feldspar are closely associated with biotite, as though formed as the biotite decomposed. These relationships are most clearly visible, although uncommon, in thin sections of the least altered gneiss in the Colton area. Much of the biotite is replaced by garnet, and most of its K has been liberated rather than trapped in K feldspar (Fig. 1).

K Feldspar in Granitized Gneiss and Gneissic Granite

An increase in K feldspar content of the gneiss is correlative with numerous field relations and textural data that indicate an addition of granite-forming substances (chiefly K) to the total rock (Engel and Engel, 1953; 1958a). Invariably the higher the K feldspar content of the rock, the higher its K content and K/Na ratio. The Ab content of associated plagioclase in the rock increases, the amount of associated biotite decreases; and the parent metasediment is altered in texture and composition toward a granite (*see especially* Engel and Engel, Part I, 1958a, Fig. 3).

The K feldspar in migmatite, granitized gneiss, pegmatite, and granite in the gneiss complex is perthitic microcline with poor to well-developed gridiron twinning. This is true throughout the region. The microcline added during granitization occurs as fine- to medium-grained replacements of plagioclase and biotite and as large porphyroblasts, anhedral to almost euhedral (Part I, Engel and Engel, 1958a, Pl. 3). Many twinned porphyroblasts either displace or replace[2] up to half a cubic inch or rarely more of the original gneiss. Most of these porphyroblasts are elongate subparallel to the foliation in the rock (Part I, Pl. 3, fig. 3). Some are clearly strained, whereas others are not, suggesting that they have grown during and slightly after deformation. Locally porphyroblasts are almost euhedral, abut against foliation planes, and enclose shreds of gneiss. Porphyroblasts of this type clearly formed by replacement (*see* Part I, Engel and Engel, 1958a, Pl. 3, fig. 2). All the K feldspar is more or less sieved with inclusions of quartz, biotite, and plagioclase. Many grains of K feldspar show Carlsbad twinning.

About 15 per cent of the rock designated and mapped as "migmatite" and "granitized gneiss" contains well-defined porphyroblasts of microcline. In the other 85 per cent of the migmatite the K feldspar added to the rock is medium- to fine-grained, so that the resulting texture is approximately equigranular. Four-fifths of the rock designated and mapped as "gneissic granite" is inequigranular principally because of the large size of the K feldspars. Most of these large K feldspars appear to be porphyroblasts, but some could have been phenocrysts in magma injected into the gneiss.

Ten specimens of K feldspar were separated from gneiss, granitized gneiss, and granite for chemical analysis. Of these, two were from Group 1A, four from Group 1, three from Group 2, and two from Group 5 (Table 1[3]).

The principal objective was to determine variations in Na, K, Ba, and in trace elements, as these are related to the physical and chemical environment of the feldspar and offer at least some information on the distribution of elements among the K feldspar, plagioclase, and other constituent minerals in the gneiss. Consequently, eight of the feldspars were analyzed for the alkali elements Na, K, Ba, and for trace elements. Two more K feldspars, one from coarse-grained granitic gneiss and one from associated pegmatite, were analyzed *in toto*. An analysis of a perthitic microcline from a pegmatite in the marble near the gneiss is included for comparative purposes.

[2] This is not to imply that the first formed K feldspar was microcline. If the initial phase was orthoclase the inversion appears to have been complete.

[3] Each specimen number of K feldspar from the paragneiss in Table 1 is given a suffix (K) to distinguish it from the host rock of the same number; that is, Qb 236 is paragneiss (rock), and Qb 236K is K feldspar from the rock. The suffix KG is added to the K feldspar from pegmatitic or granitic layers near the paragneiss of that number; that is, Qb 100KG is K feldspar from gneissic granite 65 feet from the paragneiss Qb 100 which contains the K feldspar Qb 100K. The four K feldspars from least altered gneiss Qb 228, Bgn 27, Bgn 18, and Bgn 21 are actually from slightly more feldspathic layers a few inches or a foot from the rock designated by this number. Consequently, these K feldspars (indeed, all K feldspars studied) may not have formed in equilibrium with other minerals in the indicated "host rock".

Minute traces of albite occur in several rock specimens, especially Bgn 26K, Qb 236K, Qb 100K, and Bgn 21K (Table 1), as microscopic mens of partly granitized gneiss (Bgn 26K, Qb 236K, Qb 100K) in the Emeryville-Edwards area, (2) comparisons between the microclines

TABLE 1.—COMPOSITION AND DEDUCED TEMPERATURE OF FORMATION OF PERTHITIC K FELDSPAR IN THE PARAGNEISS AND IN ASSOCIATED PEGMATITES, GNEISSIC GRANITES, AND MARBLE

Oxide or Element (Wt. per cent)	West Balmat (Group 1A)		Emeryville area (Group 1)				Edwards area (Group 2)			Colton area (Group 5)		Marble
	Bgn 26K	Bgn 27K	Qb 228K	Qb 236K	Qb 236KG1	Qb 236KG2	Qb 100K	Qb 100KG1	Qb 100KG2	Bgn 18K	Bgn 21K	P2
SiO_2	n.d	n.d	n.d	n.d	65.07	n.d	n.d	65.04	n.d	n.d	n.d	65.22
Al_2O_3	n.d	n.d	n.d	n.d	18.96	n.d	n.d	18.83	n.d	n.d	n.d	19.02
Fe_2O_3	n.d	n.d	n.d	n.d	.06	n.d	n.d	.10	n.d	n.d	n.d	.05
CaO	n.d	n.d	n.d	n.d	.16	n.d	n.d	.20	n.d	n.d	n.d	.39
Na_2O	.94	2.01	2.06	1.96	2.08	1.74	1.60	2.24	1.98	2.43	2.22	5.02
K_2O	14.62	12.75	13.32	13.22	13.18	13.28	13.38	13.08	13.30	12.82	13.04	9.12
BaO	.18	.22	.17	.25	.32	n.d	.80	.55	.27	n.d	n.d	1.07
Total					99.83			100.04				99.89
Trace Elements (In parts per million)												
B	tr	tr	tr	n.d	tr	tr	tr	tr	tr	n.d	n.d	47
Cu	5	2	4	n.d	tr	2	<2	2	4	n.d	n.d	tr
Ga	3	7	8	n.d	18	2	5	7	16	n.d	n.d	4
Mn	35	4	5	n.d	5	9	30	32	25	n.d	n.d	3
Pb	44	30	48	n.d	62	n.d	41	37	50	n.d	n.d	21
Sr	380	370	330	n.d	700	2600	1300	460	300	n.d	n.d	3700
Ti	37	41	26	n.d	14	n.d	90	50	25	n.d	n.d	16
Mol. prop.												
Ab	8.9	19.3	19.0	16.6	19.3	16.6	15.3	20.6	18.4	22.3	20.5	..
Or	91.1	80.7	81.0	81.6	80.7	83.4	84.7	79.4	81.6	77.7	79.5	..
Ab content of associated plagioclase	68	75	71	73	75	77	68	70	70	64	65	none
Temperature*	400	540	550	530	540	490	500	560	540	610	590	

Major elements, Analyst, C. G. Engel. Trace elements, Analysts, Chodos and Godijn. Looked for but not found: Ag, As, Au, Cd, Co, Cr, Ge, In, La, Mo, Nb, Pt, Sb, Sn, Ta, Th, Tl, V, W, Zn.

* Temperature of formation, in degrees Centigrade, after Barth (1956).

rims or partitions between the microcline and the plagioclase (commonly medium to sodic oligoclase). The interrelationships of these three feldspars are very similar to those in granites illustrated by Tuttle and Bowen (1958, Pl. 3), who believe the albite is unmixed from adjacent K feldspar.

The analyses of the K feldspars offer comparisons in composition of the following types: (1) comparisons of microcline in three speci- in the partly granitized gneiss and in the adjacent gneissic granite in the Emeryville-Edwards area (Qb 236K versus Qb 236KG1 and Qb 236KG2; Qb 100K versus Qb 100KG), (3) comparisons between microclines within a single body of inequigranular granite, in the same area (Qb 100KG1 and Qb 100KG2, near Edwards), (4) comparisons between K feldspars in closely associated least altered gneiss and equigranular incipiently granitized gneiss (Bgn

27 versus Bgn 26), and (5) comparisons between perthitic microcline in the same rock type but formed under very different temperatures (Bgn 26, Bgn 27, Qb 228, Qb 236, Emeryville area, compared with Bgn 18 and Bgn 21 from the Colton area). The data also indicate the approximate partitioning of Na between the K feldspar and plagioclase.

The analytical results (Table 1) show variations in composition of several microclines within a given group and rock type that are well outside analytical error. For example, very real differences in composition exist between Qb 236KG1 and Qb 236KG2 (Table 1). Differences in composition also exist between some perthitic microclines in partly granitized gneiss (Qb 100K) and that in closely associated pegmatitic granite (Qb 100KG1). There is, however, a gross correlation between Na_2O content of the perthitic microclines and (1) the appearance of albitic "rims," (2) the temperature of formation as calculated from other thermometers, and (3) the composition of associated plagioclase (Table 1).

Determinations of the Na_2O and K_2O content of the "standard" rocks G-1 and W-1, and N. B. S. feldspar No. 70 were made at the same time as the analyses of the microcline (*see* Part I, Engel and Engel, 1958a, Table 2). All the analyses are with the Perkin-Elmer flame photometer. The mean deviations recorded are much smaller than the differences in Na_2O and K_2O for most specimens in Table 1[4]. The values given for trace elements are reproducible within about 30 per cent of the number given.

Care was taken to obtain pure samples many times larger than the scale of the perthitic intergrowths. The chemical analyses include, therefore, the exsolved phase (Ab) retained within the outlines of the K feldspar grains. If, however, the perthites reflect a process of extensive unmixing and recrystallization during cooling and late-stage deformation, some of the unmixed Na, Ba, Sr, and other elements may have left some parent crystals. The migration of these elements, especially Na, would be greatly aided by deformation of the granitized rock during and immediately following feldspathization.

Kind and degree of rock deformation coincident with and continuing beyond crystal growth need not be the only, or even the prime, cause of the observed variations in composition of the K feldspars. Small but critical variations in T or P_f may have existed locally in the gneiss complex, or equilibrium may not have been attained or closely approached in all parts of the less uniformly granitized gneiss during growth of the K feldspars.

An obvious question is whether the thin albitic "screens" that separate some microcline from plagioclase are derived (unmixed ?) solely from the K feldspar, from both the feldspars, or from other sources. The question of the origin of this albite, as well as the cause of the variations in Na in the K feldspar, clouds a precise consideration of the partitioning or distribution of the alkali elements in the feldspars of these rocks. Fortunately, the amount of this albite is very small relative to the Ab content of either K feldspars or plagioclase, and it is limited to a minority of specimens. We have chosen to ignore it in subsequent calculations.

The preceding questions are especially pertinent to (1) the use of the distribution ratio of Na (as albite molecule) in K feldspar and in coexisting plagioclase as a geologic thermometer (Barth, 1956; Yoder *et al.*, 1957, p. 206–214; Taylor and Heier, 1958), (2) the calculation of distribution coefficients or factors (Eugster, 1955; Taylor and Heier, 1958), and (3) the argument that the large crystals of K feldspar replacing paragneiss adjacent to "granite" have properties identical to K feldspars in the granite (Raguin, 1946, p. 183; Read, 1948, p. 12–14; Perrin, 1956).

In attempting to use the distribution of Na between coexisting K feldspar and plagioclase as a geologic thermometer, Barth (1956, p. 1–16) has deduced an "average curve" indicative of the temperatures of crystallization. A serious complication recognized by Barth is the lack of experimental data on the various structural inversions of K feldspar. It is also necessary that the plagioclase and K feldspar grow in equilibrium and that the initial ratio of Na in the feldspars during their growth remain fixed in the crystals after cooling and aging. The presence of albite rims of ambiguous origin in several specimens of the granitized Adirondack paragneiss is an especially complicating factor and a probable source of error. There is no such albite described in the rocks that Barth (1956) and Taylor and Heier (1958) studied; nor do these workers appear to correct for pressures or for Ba and other "trace" alkalis common in many of the K feldspars in paragneisses and granitic gneisses[5]. Analyses of K

[4] There are no established differences in Na_2O and K_2O between specimens Qb 228K and Qb 236K or between Qb 100KG1 and Bgn 21K.

[5] After this was written and presented for publication, Heier and Taylor published their data on the distribution of Li, Na, K, Rb, Cs, Pb, and Tl in

feldspars in Table 1 indicate amounts of BaO reaching 0.8 weight per cent. The problems cited above seem sufficiently complex to inhibit the use of our analytical data on the feldspars in geologic thermometry.

We have, however, calculated the temperatures of formation of the feldspars, by use of Barth's deduced curve (1956, Fig. 9). The temperatures of crystal growth so calculated are, for the Emeryville area, ∽ 500°–550° C, and for the Colton area ∽ 600°–610° C (Table 1). These figures, whatever their validity, are essentially those derived from other thermometers (Part I, Engel and Engel, 1958a) and in the range of temperatures suggested for the metamorphism of many gneisses of southern Norway (Barth, 1956; Taylor and Heier, 1958).

The statement that porphyroblastic K feldspars in paragneiss are "identical" to the K feldspars in immediately associated granites finds its typical expression in the work of Perrin who notes (1956, p. 3):

"... A fundamental fact, observed by many others before me, in particular by such eminent French authors as Auguste Michel Levy, Ch. Barrois, Ch. Lacroix, and by many others since then, namely the development in the country rock (mica schists for instance) near the granite contact, of feldspar crystals thoroughly identical with those of the granite in all their features: size, shape, *analysis* [present authors' italics], color, inclusion of relicts of biotite or hornblende, etc.".

Several other disciples of the "soak" school have made equally strong statements and offered this purported fact as evidence of the metasomatic origin of the granite (Raguin, 1946, p. 183; Read, 1948, p. 12–14). All these statements are characterized by the absence of supporting data. All are made to prove that the granite is metasomatic. Actually it seems possible that nature permits the evolution of a rock system involving metasediment, invading granitic magma, and fluids therefrom in which the claims of Perrin might be approached or achieved. The similarity of K feldspars in granite and enclosing gneiss is *not* proof that the granite is metasomatic. But is seems a worthwhile prerequisite actually to determine to what extent porphyroblastic K feldspars in most or much gneiss enveloping "granite", and in the granite, are "identical" even by existing techniques of measurement.

The pair of specimens Qb 100K and Qb 100KG1 are of particular interest with respect to the statements of Perrin, Raguin, Read, and other advocates of the occurrence of "identical" K feldspars. Specimen Qb 100K is an augen gneiss in which the "eyes" of microcline are subhedral, up to an inch long, and typical of the porphyroblastic K feldspar from little deformed migmatite and granitized gneiss. Analogous porphyroblastic gneisses are described in granitized terranes throughout the world. Some of the microcline porphyroblasts abut against the foliation. Detailed mapping shows these are related in space and presumably in origin to coarse microclines in the nearby pegmatitic granite, Qb 100KG1 and Qb 100KG2 (Fig. 3). To the field observer the microclines in Qb 100K, Qb 100KG1, and Qb 100KG2 do "look alike in all respects"; but the fact that these feldspars are not compositionally alike is hardly surprising for reasons noted above. Considering their origin in a system complicated by gradients in composition, by deformation during cooling, and possibly by local, although small, variations in T and P it would be rather surprising if they were identical.

Plagioclase Feldspar

Introduction

Studies of plagioclase in the gneiss involve Xrays (powder diffraction), analyses of trace elements, and the microscopic examination necessary to determine composition, form, twinning, and degree of alteration. Eight plagioclases were X-rayed (Qb 228, Qb 230, Qb 231, Bgn 18, Bgn 19, Bgn 20, Bgn 21, Bgn 27). About 200 thin sections containing plagioclase were examined optically. Trace-element analyses were made of 8 specimens. Compositions of 29 plagioclases were determined by conventional oil-immersion techniques. These determinations of composition from optical data were double-checked in the specimens where we have chemical analyses of total rock and associated minerals. Probably the plagioclase determinations are accurate to ±1.5 Ab.

Analytical Methods

The trace-element analyses of plagioclase were obtained with the same equipment and by the same methods described in Part I (Engel and Engel, 1958a, p. 1378–1379). Sensitivities and reproducibilities are as previously reported.

X-ray powder patterns were taken with a

southern Norwegian Precambrian alkali feldspars (1959). Many of the metamorphic felspars examined by Heier and Taylor and by Barth (1956) have compositions very close to specimens in Table 1.

Norelco high-angle diffractometer using copper K α radiation. Powder mounts were made by placing the powdered plagioclase in a shallow aluminum cell according to the method described by Yoder and Eugster (1954, p. 162). The reflections chosen for study were ($1\bar{3}1$) and (131) for the reasons Smith and Yoder (1956a) have noted: (1) they can be clearly resolved from neighboring reflections, (2) their 2θ values vary sensitively with composition, and (3) the reflections are close enough so that measurements of them minimize chart-scale errors.

The appropriate 2θ region was scanned three times. Measurements were averaged and show a maximum deviation from the mean of 0.01° 2θ. The precision of the average value of a reflection separation for any sample is probably $\pm$0.01° 2θ.

TABLE 2.—CONTRASTING PROPERTIES OF PLAGIOCLASE IN LEAST ALTERED GNEISS, EMERYVILLE AND COLTON AREAS, NEW YORK

Property	Emeryville area (Group 1) Max T 500° C	Colton area (Group 5) Max T 600° C
Volume per cent in rock	$\bar{x}$ = 40	$\bar{x}$ = 49
	Range Ab 32–Ab 47	Range Ab 42–Ab 55
Composition	$\bar{x}$ = Ab 73	$\bar{x}$ = Ab 67
	Range Ab 88–Ab 38	Range Ab 74–Ab 46
Twinning (as viewed in thin section)	5% grains visibly twinned	95% grains visibly twinned
	Twin types = pericline and albite	Twin types = albite; pericline rare
Grain size (mean grain diameters)	0.18 mm	0.55 mm
Alteration	Moderate, largely sericitization (10% ? feldspar replaced by sericite)	Slight, principally sericitization (5% ? feldspar replaced by sericite)
Zoning	Very rare	Very rare

$\bar{x}$ = average composition

Plagioclase in Least Altered Gneiss

The average plagioclase in the least altered gneiss at Colton is more abundant, more calcic, coarser-grained, less altered, and more clearly and pervasively twinned than that at Emeryville. X-ray studies suggest it is markedly inverted toward an as yet undefined low-temperature form. Most of these relationships of plagioclase in least altered gneiss are summarized in Table 2. Expanded data on the range in composition of these plagioclases are in Table 3. Optical studies of plagioclase in least altered gneiss at points intermediate between Emeryville and Colton indicate a transition in composition between the extreme types. Statistical comparisons of about 50 plagioclases found along the strike of least altered gneiss from Emeryville to Edwards (12 miles) indicate the following changes: (1) a decrease in Ab content of the mean plagioclase from Ab 73 to Ab 71; (2) an increase in the abundance and clarity of polysynthetic twinning (5 per cent of the grains at Emeryville show twinning, whereas 25 per cent are visibly twinned at Edwards); (3) a 50 per cent increase in mean diameter of plagioclase grains. Data on the plagioclase in the least altered gneiss at Russell are less extensive, but the mean composition is about Ab 69. Eighty per cent of the grains obviously are twinned, and the mean diameter of grains is more than two times that at Emeryville. At Colton the mean plagioclase has the composition Ab 67 and a mean diameter more than three times that of the plagioclase at Emeryville.

The brief summary in Table 3 indicates the increase in An content of the plagioclase that accompanies the increase in "grade" or "rank" of metamorphism. The relations are consistent with the generalization that in argillaceous sandstones metamorphosed to or beyond the upper amphibolite facies (in the classical terminology)[6] oligoclase is the characteristic plagioclase. The composition of the plagioclase

[6] Probably there is a sequence of specific mineral "isograds" depending on the bulk composition of the system; hence the correlations in Table 4 are gross at best.

can be influenced strongly by the composition of the system as is apparent not only from our data but from those of Harker (1932), Barth (1936), Billings (1937), and many others.

Plagioclase of Ab 75 to Ab 68 appears to be typical of the "Sillimanite zone" as mapped in the Wissahickon schist of Pennsylvania (Weiss, 1949), in Dutchess County, New York (Barth, 1936), and in the Dalradian of Scotland (Phillips, 1930). This composition (Ab 75 to Ab 68) is about that of the mean plagioclase in the paragneiss of the Emeryville area where sillimanite has formed in associated pelitic units that are higher than the paragneiss in Al and K and lower in Ca (Brown and Engel, 1956, p. 1604). In the Adirondack paragneiss, and in most other gneisses and schists from which data are available, the increase in An content of plagioclase with increasing grade of metamorphism is accompanied by a decrease in its range of composition. Presumably, the relations indicate a closer approach of the system toward equilibrium as temperature of metamorphism increases.

TABLE 3.—COMPOSITIONS OF PLAGIOCLASE IN LEAST ALTERED GNEISS FROM THE EMERYVILLE AND COLTON AREAS (Groups 1 and 5), NEW YORK

Area and group	Ab 80–Ab 90	Ab 70–Ab 80	Ab 60–Ab 70	Ab 50–Ab 60	Ab 40–Ab 50	Ab 30–Ab 40
	Number of specimens					
Emeryville area (Group 1)	1	21	11	5	2	1
Colton area (Group 5)		11	16	3	2	..

Most of the alteration in the plagioclase in least altered gneiss is very fine-grained sericite. The sericite occurs in irregular patches, in zones along some of the twin lamellae, and along visible fractures and other physical imperfections in the gneiss. It is only rarely disseminated uniformly through the plagioclase. Sericitization appears to have been a retrograde process, and in the Emeryville area the sericite is clearly later than, and easily distinguished from, the much coarser-grained muscovite in the rock. Possibly the sericite was produced by leaching of K from the plagioclase itself. But if the K in the sericite was introduced into the least altered gneiss at Emeryville during retrograde metamorphism it follows that the differences in K induced in the gneiss between this area and Colton during the apex of metamorphism (Fig. 1)[7] were reduced during the waning stages.

The great increase in polysynthetic twinning and the change in the twin types in the plagioclase between Emeryville and Colton are striking features of the least altered gneiss. Possibly pericline twinning is as common as, or commoner than, albite twinning in the plagioclase at Emeryville although the observations are too few to prove this (Table 2). In contrast, pericline twinning appears uncommon in the Colton area, whereas albite twin lamellae appear in almost every grain. The large increase in albite twinning at higher grades of metamorphism is accompanied by a broadening in width of the twin lamellae. Abundant pericline twinning exists in oligoclase of the metamorphosed "Green Beds" of the Scottish Dalradian (Phillips, 1930), and possibly pericline twinning is a common or dominant (?) feature of many metamorphic plagioclases in the compositional range Ab 70–Ab 80.

The absence of zoning in the plagioclases in the gneiss is in sharp contrast to the zoned grains described from the Wissahickon (Weiss, 1949, p. 1692), the Green Beds of the Dalradian, the Paleozoic metasedimentary rocks of northeastern Japan (Sugi, 1935), and elsewhere (Table 4).

Published data on the variations in the kind and amount of plagioclase in pelitic rocks, as a function of progressive metamorphism, also are summarized in Table 4. There is little specific comment, however, on the amount and degree of alteration (sericitization) of plagioclases in the middle and upper zones of metamorphism. The implied relationships seem to be those we observe: that the oligoclase and oligoandesines of the middle grade or lower high-grade schists and gneisses are in general more sericitized than the more calcic plagioclases found in extremely high-grade metamorphic rocks. Our collection of Alpine paragneisses show similar features, and C. E. Tilley notes (oral communication, 1958) that this relation holds for the plagioclase in some of the metamorphosed clastic rocks of the Scottish Highlands.

X-Ray Powder Diffraction Data

The angular separation between the ($1\bar{3}1$) and (131) reflections in X-ray diffractometer

[7] By basification, as discussed in Engel and Engel, 1958a, Part 1, p. 1397–1399.

TABLE 4.—PLAGIOCLASE FELDSPAR IN METAMORPHOSED CLASTIC SEDIMENTS OF DIVERSE AGES AND LOCALITIES

Locality and age of rocks	Metamorphic rank or zone					Zoning	Reference
	Biotite	Garnet	Staurolite	Kyanite	Sillimanite		
Dutchess County, New York (Paleozoic)	Ab 95–Ab 90	Ab 95–Ab 90	Ab 95–Ab 65	Ab 95–Ab 50	$\bar{x}$ = Ab 70	Uncommon	Barth (1936)
New Hampshire (Paleozoic)	$\bar{x}$ = Ab 90 Ab 100–Ab 80	? Ab 90				?	Billings (1937)
Scotland (Precambrian)		$\bar{x}$ = Ab 93 Ab 95–Ab 90		$\bar{x}$ = Ab 68 Ab 90–Ab 63			Phillips (1930)
Pennsylvania (Paleozoic ?)		$\bar{x}$ = Ab 85 Ab 100–Ab 70		$\bar{x}$ = Ab 85 Ab 100–Ab 60	$\bar{x}$ = Ab 75 Ab 100–Ab 60	Common	Weiss (1949)
South Westland, New Zealand (Mesozoic)	$\bar{x}$ = Ab 93	Oligoclase					Turner (1933)
Northern Michigan (Precambrian)		Oligoclase (~ Ab 85)					James (1955)
Abukuma Plateau, Japan (Paleozoic)	$\bar{x}$ = Ab 90 Ab 95–Ab 85	Oligoclase-Andesine				Common	Miyashiro (1953; 1958) Sugi (1935)

$\bar{x}$ = average composition

TABLE 5.—X-RAY POWDER DIFFRACTION DATA ON PLAGIOCLASE FELDSPARS IN LEAST ALTERED GNEISS, EMERYVILLE AND COLTON AREAS, NEW YORK

Area	Sample	Reflections*		Mean 2θ separations*
		2θ (131)	2θ ($1\bar{3}1$)	
Emeryville	Bgn 27	31.39	29.77	1.62
(Group 1)	Qb 230	31.36	29.77	1.59
	Qb 231	31.36	29.77	1.59
	Qb 228	31.38	29.76	1.62
	$\bar{x}$	31.37	29.77	1.61
Colton	Bgn 18	31.42	29.72	1.70
(Group 5)	Bgn 20	31.40	29.71	1.69
	Bgn 21	31.42	29.71	1.71
	Bgn 19	31.42	29.70	1.72
	$\bar{x}$	31.42	29.71	1.71

* Each reading of 2θ (131) and 2θ ($1\bar{3}1$) reported here is the mean of three exposures of the same mount in which the maximum mean deviation is .007. Use of separate mounts of the same sample induced a maximum range of 0.07°, 2θ and in other samples the range is 0.04° or less. Probably the accuracy of the mean value of a reflection separation for any sample is ± 0.02°θ.

patterns of eight plagioclases from least altered gneiss are plotted in Table 5. The data also are plotted against composition in Figure 4, which employs curves developed by Smith and Yoder (1956a). These authors conclude that the curves—defined by synthetic and igneous plagioclases—indicate (1) the degree of inversion of plagioclase from a "high" to some undefined low-temperature state and (2) the difficulties (impossibility ?) of setting up generalized curves from powder-diffraction patterns that will yield composition of plagioclases of diverse origin.

The plagioclases from the paragneiss lie slightly below the composition—angular separation curve defined by plagioclases from thick stratiform mafic intrusives, but along or close to the curve defined by plagioclases from granites and pegmatites (Fig. 4).

BIOTITE

Introduction

The properties of biotites vary markedly in the paragneiss, especially when comparisons are made between Emeryville and Colton. Smaller variations exist among biotites of any single group, but these in-group variations do

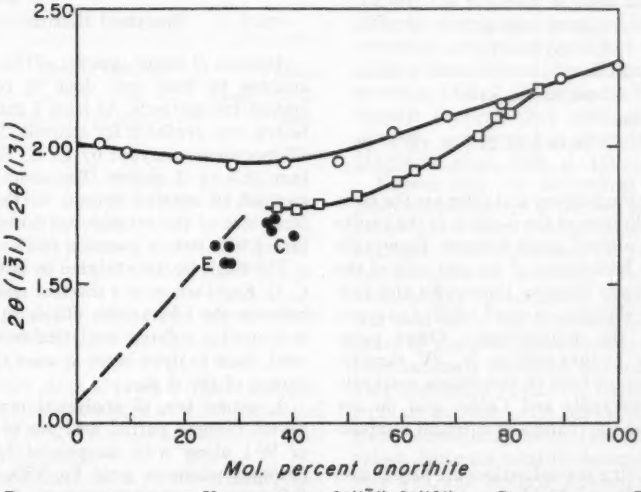

FIGURE 4.—RELATIONSHIPS OF THE VARIATIONS IN $2\theta(1\bar{3}1)$–$2\theta(131)$ TO COMPOSITION IN PLAGIOCLASES IN THE PARAGNEISS COMPARED WITH OTHER PLAGIOCLASES

Solid circles designated *E* are plagioclases from Emeryville; solid circles designated *C* are plagioclases from Colton. Open squares are plagioclases from thick stratiform mafic intrusions, open circles are plagioclases synthesized the dry way, and dashed line is curve defined by plagioclases from granites and pegmatites (after Smith and Yoder, 1956a).

not appear to be closely related to differences in bulk composition of the paragneiss. Except for differences in color, biotites from least altered gneiss (Group 1) are not distinguishable from biotites in incipiently granitized gneiss.

TABLE 6.—COMPOSITION OF BIOTITE SPECIMEN Qb 100 AS OBTAINED BY DIFFERENT ANALYSTS

Oxide	Analyst		Mean	Mean deviation
	L. C. Peck	C. G. Engel		
SiO_2	35.60	35.82	35.71	.11
TiO_2	4.18	4.27	4.22	.05
Al_2O_3	15.76	15.91	15.84	.08
Fe_2O_3	1.19	2.57	1.88	.69
FeO	21.51	19.48	20.50	1.02
MnO	.22	.26	.24	.02
MgO	8.18	8.24	8.21	.03
CaO	.34	.37	.36	.02
Na_2O	.18	.22	.20	.02
K_2O	9.00	8.99	8.99	.01
H_2O^+	3.05	3.10	3.08	.03
H_2O^-	.11	.08	.10	.02
P_2O_5	n.d.	.09	..	..
F	.47	.41	.44	.03
	99.79	99.81		
Less O — F	.20	.18		
Total	99.59	99.63		
Total Fe (as Fe_2O_3)	24.09	24.21	24.15	.06

Chemical composition and color are the most sensitive indicators of the position of the biotite in the least altered gneiss between Emeryville and Colton. Dimensions of the unit cells of the biotites decrease between Emeryville and Colton, but the changes are small relative to errors inherent in the measurements. Other properties of the biotites such as N_z, 2V, density, and polymorphic form do not change markedly between Emeryville and Colton and do not clearly reflect the changes in chemical composition.

The preceding generalizations are based upon detailed studies of 35 biotites from the paragneiss. Twenty-four biotites were analyzed for major, minor, and trace elements. Lee C. Peck made 8 of the analyses of major elements; C. G. Engel analyzed 1 of these samples and 16 additional biotites. Partial analyses were made on 12 other biotites. These partial analyses commonly include determinations for FeO, MnO, TiO_2 and trace elements. The optical value N_z was determined on 32 biotites. Densities were measured on 8, and X-ray powder-diffraction measurements were made on 12 biotites. A single-crystal (Weissenberg) study was made on 1 biotite. We are deeply indebted to A. A. Levinson, B. Raychaudhuri, and H. S. Yoder for assistance in the X-ray studies.

TABLE 7.—COMPARISONS OF "RAPID" AND CONVENTIONAL ANALYSES FOR TOTAL IRON (as Fe_2O_3) IN SPECIMENS OF BIOTITE Analyst, C. G. Engel

Specimen	Type and date of analysis			Mean	Mean deviation
	Rapid* .2000 gm 5/16/55	Rapid* .1000 gm 9/3/55	Conventional† 7/9/55		
Bgn 4	21.62	21.90	21.69	21.74	.12
Bgn 9	17.97	18.20	18.37	18.18	.14
Qb 226	18.55	19.04	18.63	18.74	.20

* See Shapiro and Brannock (1956, p. 35–36)
† Using silver reductor

Analytical Methods

Analyses of major elements.—The 8 chemical analyses by Peck were done by conventional gravimetric methods. At least 4 grams of each biotite was available for analysis. Most of the 17 specimens analyzed by C. G. Engel varied from 1.7 to 4 grams. The modifications in methods of analysis because of the relatively small size of the samples are noted below together with tests of precision and accuracy.

The single biotite analyzed by both Peck and C. G. Engel serves as a check of reproducibility between the laboratories (Table 6). Inasmuch as somewhat different analytical techiques were used, these analyses serve as some check on accuracy of the data.

A second test of analytical error was obtained through partial analyses of either G-1 or W-1 along with samples of biotites. For example, whenever total Fe, TiO_2, MnO, and P_2O_5 were run on the spectrophotometer, a sample of W-1 also was run. The choice of W-1 or G-1 was based upon which of these rocks had concentrations of the desired elements closest to the element sought in the biotite.

Obviously a "standard" biotite would have been far superior to either G-1 or W-1, and perhaps before another decade passes standards for several minerals can be set up.

TABLE 8.—REPRODUCIBILITY OF FeO DETERMINATIONS IN BIOTITES BY DIFFERENT ANALYSTS

Specimen	Analyst		Difference	Mean	Mean deviation
	L. C. Peck	C. G. Engel and A. E. Engel			
Qb 100....	21.51	19.48*	2.03	20.50	1.015
Qb 229....	19.06	18.75	0.31	18.91	.155
Qb 3.....	18.36	17.87	0.49	18.12	.245
Qb 235M..	17.71	17.35†	0.36	17.53	.180
Qb 236....	16.65	16.32	0.33	16.49	.165

* Average of two analyses (19.50, 19.45 per cent FeO)

† .2500 gm. sample. Other analyses on .5000 gm sample

TABLE 9.—ANALYSES OF FeO IN REPLICATE SAMPLES OF BIOTITE MADE AT INTERVALS OF MONTHS TO TEST PRECISION OF ANALYTICAL METHODS

Analysts, A. E. Engel and C. G. Engel

Sample	Date of analysis			Mean	Mean deviation
	5/24/55	7/16/55	9/12/55		
Bgn 20..	14.36	14.61	14.72	14.56	.14
Bgn 18..	13.53	13.76	13.70	13.66	.09
Bgn 25..	16.25	16.30	16.39	16.31	.05
Bgn 6...	17.72	17.66	17.95*	17.78	.13
Qb 7....	18.48	18.39	18.76	18.54	.14

* .2500 gm sample; other values determined from .5000 gm samples

As another check of precision, identical splits of biotites were reanalyzed at widely separated times for Fe_2O_3, FeO, TiO_2, Na_2O, and K_2O (Tables 7, 9, 10, 12).

An independent check of the accuracy and precision of analyses of Mn was obtained by comparison of the wet chemical and spectrochemical values (Table 11).

Whenever the size of the sample permitted, the main portion was done on 1 gram of the mineral. In 5 instances, SiO_2, Al_2O_3, CaO, and MgO were determined on a 0.5-gram sample. Total Fe was obtained with a silver reductor on most specimens, but in 12 samples (*see* Tables 7 and 13) total Fe was obtained on the spectrophotometer using either a 0.2000- or 0.1000-gram sample as permitted by sample size (Shapiro and Brannock, 1956, p. 23–24, 35–36). To test the precision of these analyses, where different sample sizes were involved, both conventional and "rapid" analyses were made of splits of three biotites (Bgn 4, Bgn 9, Qb 226), as shown in Table 7. The reproducibility obtained appears to equal that obtained by different analysts employing conventional means (Fairbairn *et al.*, 1951, p. 15).

TABLE 10.—REPLICATE ANALYSES OF TiO_2 IN BIOTITES, DONE AT INTERVALS OF MONTHS, AS A GUIDE TO PRECISION OF THE ANALYTICAL METHODS

Analyst, C. G. Engel

Specimen	Weight of sample and date analyzed			Mean	Mean deviation
	0.100 gm 5/7/55	0.100 gm 7/8/55	0.200 gm 1955–1956		
Bgn 18..	5.01	5.08	4.97*	5.02	.04
Bgn 21..	5.48	5.68	5.55*	5.57	.07
Bgn 9...	4.12	4.19	3.94*	4.08	.10
Qb 227..	3.21	3.23	3.31*	3.25	.04

* Value obtained during complete analysis of biotite sample and cited in Table 13

Ferrous iron was determined on a 0.5000-gram sample, except in the two instances designated where only a 0.2500-gram sample was available. Determinations of FeO employed the conventional technique (Fairbairn *et al.*, 1951, p. 15). An abnormal amount of biotite was used up in the determinations largely because of the differences between the FeO determinations in biotite Qb 100 analyzed by both Peck and Engel (Table 6). Inspection of these replicate analyses of Qb 100 show that the only major discrepancy is in the FeO and Fe_2O_3 values. Since the total Fe determinations agree very well, and Fe_2O_3 is obtained by difference, the lack of reproducibility is in FeO.

As a further check between the FeO values obtained by Peck and by Engel, 5 biotites analyzed by Peck were reanalyzed for FeO by C. Engel and A. Engel (Table 8). All our values

TABLE 11.—COMPARISON OF THE Mn CONTENT OF IDENTICAL BIOTITES AS DETERMINED BY SPECTROPHOTOMETER AND EMISSION SPECTROGRAPH

Area and group	Sample	Spectrophotometer weight per cent MnO	Emission spectrograph	
			Weight per cent MnO	Parts per million Mn
Emeryville	Bgn 26	.25	.26	2000
(Group 1)	Bgn 27	.17	.26	2000
	Qb 236	.28	.19	1500
	Qb 235M	.30	.19	1500
	Qb 3	.30	.26	2000
	Qb 7	.16	.12	1000
	Qb 230	.29	.21	1600
	Qb 229	.14	.10	800
	Qb 231	.22	.14	1125
	Qb 226	.23	.18	1400
	Qb 227	.25	.21	1600
	Qb 228	.14	.08	650
Edwards	Qb 233	tr	..	140
(Group 2)	Qb 100	.22	.19	1500
	Qb 216	.04	.03	250
Russell	Bgn 4	tr	..	100
(Group 3)	Bgn 6	tr	..	150
	Bgn 9	.04	.03	250
	Bgn 14	.03	.03	240
Colton	Bgn 23	.07	.09	500
(Group 5)	Bgn 24	tr	..	100
	Bgn 18	tr	..	100
	Bgn 20	tr	..	100
	Bgn 21	tr	..	80
	Bgn 19	..	..	60

Spectrophotometer analyses by C. G. Engel
Spectrographic analyses by Chodos and Godijn

are lower than Peck's by roughly 3 per cent of the number reported except Qb 100 where our determination is almost 10 per cent below his. Inasmuch as the analyses were made on splits of the same sample, a further check on the reproducibility of our own analyses seemed desirable. Table 9 summarizes results of 3 separate FeO determinations on each of 5 biotite specimens where material was available. The reproducibility indicated by these analyses presumably applies to all of our own analyses of FeO, but this means our FeO values are consistently lower than Peck's by at least 2 per cent of the value reported.

As noted previously, the rock W-1 (Fairbairn *et al.*, 1951) was analyzed for FeO along with 2 separate groups of biotites. The values of FeO for W-1 obtained by us are listed in Part I, Table 1, along with the revised mean values of FeO as calculated by Fairbairn. No large discrepancies exist in these data.

The amounts of TiO_2 in the 15 biotites analyzed by C. Engel were determined on the Beckman Model B spectrophotometer. At least 2 TiO_2 determinations were made on most specimens at intervals of days or weeks. In 1 of the analyses a 0.2000-gram sample was used; the second employed a 0.1000-gram sample. Each of 4 specimens was analyzed 3 times to obtain a further measure of precision. The analyses in triplicate are listed in Table 10, along with means and mean deviation. Although the 0.1000-gram samples were used in part because of scarcity of sample, the high TiO_2 content of the biotites actually required dilution before it could be run on the spectrophotometer.

TABLE 12.—REPRODUCIBILITY OF K_2O AND Na_2O IN IDENTICAL SAMPLES OF BIOTITES USING FLAME PHOTOMETER AND J. LAWRENCE SMITH ANALYTICAL PROCEDURES

Analyst, C. G. Engel

Sample number	K_2O			Na_2O		
	J. L. Smith	Flame photo.	Mean deviation	J. L. Smith	Flame photo.	Mean deviation
Bgn 25..	9.02*	8.76	.13	.40*	.52	.06
Bgn 4 ..	7.96*	7.67	.15	.37*	.31	.03
Bgn 9 ..	8.23*	8.01	.11	.58*	.52	.03
Qb 233..	8.89*	8.81	.04	.21*	.32	.06

* The value listed in Table 13 and used in calculations

Comparisons of amounts of TiO_2 in biotites obtained from emission spectroscopic analyses compared with amounts recorded by spectrophotometer show no large discrepancies.

MnO was determined on an aliquot of the same solution used to determine TiO_2. In addition, Mn was determined by Chodos and Godijn with the emission spectrograph. The wet chemical and spectrographic values for MnO are listed in Table 11. The same trends are apparent in both types of analyses, although the spectrographic values are generally lower than the values obtained on the spectrophotometer. On the basis of this analytical work the lower limit of Mn determinable on the spectrophotometer is in the range 100–150 parts per million.

TABLE 13.—CHEMICAL ANALYSES, N_D, COLOR, AND DENSITY OF BIOTITES FROM THE PARAGNEISS IN THE EMERYVILLE-COLTON REGION, NEW YORK

Wet Balmat	Emeryville Group 1	Edwards Group 2	Russell Group 3	South Pierrepont	Colton Group 5

TABLE 13.—CHEMICAL ANALYSES, N_g, COLOR, AND DENSITY OF BIOTITES FROM THE PARAGNEISS IN THE EMERYVILLE-COLTON REGION, NEW YORK

	West Balmat Group 1A		Emeryville Group 1										Edwards Group 2			Russell Group 3				South Pierrepont Group 4	Colton Group 5		
	Bgn 26	Bgn 27	Qb 236	Qb 235M	Qb 3	Qb 7	Qb 230	Qb 229	Qb 231	Qb 226	Qb 227	Qb 228	Qb 233	Qb 100	Qb 216	Bgn 6	Bgn 9	Bgn 14	Bgn 4	Bgn 25	Bgn 18	Bgn 21	Bgn 19
SiO_2	37.23	35.42	36.61	36.16	35.02	35.38	35.32	35.41	35.38	36.70	35.72	35.27	36.10	35.60	35.37	36.62	37.59	36.96	36.63	35.18	37.97	36.24	36.58
TiO_2	3.27	3.15	3.01	3.18	3.11	3.20	2.97	2.84	3.40	2.92	3.31	3.43	4.11	4.18	4.33	4.58	3.94	4.79	4.25	3.94	4.97	5.55	5.14
Al_2O_3	18.26	19.04	17.41	17.40	17.78	20.38	18.94	18.74	18.25	19.04	19.79	18.66	18.48	15.76	17.50	17.55	18.84	16.94	17.03	17.90	17.54	17.24	17.35
Fe_2O_3	1.44	2.70	2.30	2.55	3.00	1.50	1.69	1.61	2.34	1.89	1.89	1.36	2.42	1.19	1.41	1.17	.71	.96	1.02	1.62	1.63	1.14	1.13
FeO	18.69	16.11	16.65	17.71	18.36	18.54	18.85	19.06	19.89	15.17	17.46	19.22	16.29	21.51	19.25	17.78	15.73	18.72	18.65	16.32	13.66	15.30	13.93
MnO	.07	.25	.28	.07	.30	.16	.29	.14	.22	.23	.25	.14	.00	.22	.04	.00	tr	tr	.00	.00	.00	.00	.00
MgO	8.25	9.56	10.37	9.36	9.10	7.77	9.11	8.92	7.33	10.79	8.93	8.60	9.61	8.18	9.12	9.79	9.80	8.59	9.75	12.01	11.22	11.12	13.08
CaO	.10	.24	.04	.30	.09	.02	.02	.00	.03	.34	.16	.13	.28	.34	.24	.26	.80	.44	.20	.07	.90	.76	.18
Na_2O	.20	.40	.09	.12	.36	.41	.36	.19	.52	.36	.42	.08	.21	.18	.10	.38	.58	.51	.37	.40	.38	.42	.07
K_2O	9.28	9.30	9.64	9.61	8.67	8.93	8.86	9.33	9.14	8.86	8.09	9.34	8.89	9.00	9.09	8.32	8.23	8.32	7.96	9.02	8.20	8.70	8.73
H_2O+	3.16	3.48	3.13	3.30	3.93	3.46	3.39	3.52	3.09	3.41	3.54	3.44	3.27	3.16	3.28	3.48	3.55	3.30	3.92	3.41	3.09	3.28	3.38
F	.25	.37	.31	.29	.21	.24	.37	.43	.43	.27	.27	.33	.32	.47	.26	.40	.35	.27	.30	.37	.39	.43	.52
	100.20	100.02	99.84	100.05	99.93	99.99	100.17	100.19	100.02	99.98	99.83	100.00	99.98	99.79	99.99	100.33	100.12	99.80	100.08	100.24	99.95	100.18	100.09
O − F	.11	.16	.13	.12	.09	.10	.16	.18	.18	.12	.12	.14	.14	.20	.11	.17	.15	.12	.13	.16	.17	.18	.22
Total	100.09	99.96	99.71	99.93	99.84	99.89	100.01	100.01	99.84	99.86	99.71	99.86	99.84	99.59	99.88	100.16	99.97	99.68	99.95	100.08	99.78	100.00	99.87
Trace Elements (In parts per million)																							
Ba	300	750	700	500	400	540	625	500	620	760	500	750	550	1000	600	500	350	1800	900		1000	2200	1800
Co	40	40	35	40	45	47	44	35	42	30	45	30	56	35	30	25	40	53	35		50	55	55
Cr	150	130	100	100	150	200	210	100	140	130	180	150	210	150	100	200	150	240	100		200	300	350
Cu	10	9	10	9	50	25	25	30	19	85	30	25	160	40	30	100	20	28	30		100	25	30
Ga	40	48	40	45	50	35	40	55	72	50	50	52	57	40	55	70	60	35	45		60	50	80
Mn	2000	2000	1500	1500	2000	1000	1600	800	1100	1400	1600	650	140	1500	250	150	250	240	100		100	80	60
Nb	20	35	30	20	10	38	30	30	100	27	37	30	50	20	30	50	30	30	30		50	20	30
Ni	80	80	70	70	100	110	85	70	83	45	85	50	110	40	70	60	100	120	70		100	120	120
Sc	70	66	70	70	70	60	68	70	85	54	58	80	42	60	60	40	50	38	60		20	22	15
Sr	15	20	15	15	15	20	20	15	17	20	20	15	20	15	15	15	15	20	15		20	20	15
V	300	320	200	230	300	390	430	300	400	410	350	300	480	300	400	350	500	500	400		500	600	700
Y	20	27	30	30	20	31	20	40	25	28	24	45	35	20	20	25	50	15	20		30	25	10
Yb	4	4	3	3	5	4	3	5	3	3	3	4	3	5	4	4	5	2	4		3	2	2
Zr	60	80	40	40	50	120	80	100	120	75	80	75	130	30	100	100	100	115	80		200	120	120
N_g	1.644	1.640	1.641	1.647	1.646	1.647	1.644	1.645	1.647	1.635	1.644	1.644	1.644	1.648	1.646	1.645	1.639	1.649	1.647	1.639	1.643	1.644	1.641
Color (Y)*	gbl	gbl	gbr	gbr	gbr	gbr	rbr	br	gbr	rbr	gbr	br	rbr	rbr	rbr	drbr	drbr	drbr	dr	gbr	rbl	drbr	drbr
Density		3.147		3.146	3.150	3.143								3.148							3.155	3.148	3.152

Elements looked for but not found: Ag, As, Au, Bi, Cd, Ge, In, La, Mo, Pt, Sb, Sn, Ta, Th, U, Zn

* gbl = greenish black, gbr = greenish brown, rbr = reddish brown, br = brown, drbr = deep reddish brown, dr = deep red, rbl = reddish black

Qb 235M, Qb 236, Qb 216, Qb 228, Qb 229, Qb 100, Qb 3, Qb 231; Lee C. Peck, analyst

Qb 7, Qb 230, Qb 227, Qb 233, Bgn 26, Bgn 27, Bgn 6, Bgn 9, Bgn 14, Bgn 4, Bgn 25, Bgn 18, Bgn 21, Bgn 19; Celeste G. Engel, analyst

Values for K_2O and Na_2O in the biotites were determined by the J. L. Smith method when ever 0.500 gram of sample was available. Where only 0.2000 gram of biotite could be spared alkalis were determined with the Beckman Model DU spectrophotometer with flame attachment. As a measure of precision between the two methods, each of 4 specimens was analyzed both ways, at different times during 2 years. The results of these analyses are given in Table 12. These data suggest that the K_2O values obtained in this study using the flame photometer are consistently lower than those obtained by the J. L. Smith method, by as much as 3 per cent of the value obtained. We have not attempted a correction of the flame-photometer results but have designated in all tables the analyses of K and Na by flame photometer.

All values for water were obtained from a 0.500-gram sample fused with lead oxide in a furnace heated initially to 850° C.

The 8 biotites analyzed by Peck for F were used as standards with which to obtain F spectrographically in the other biotites. The spectrographic analyses are reproducible to within about 15 per cent of the amounts reported.

Determinations of optics, color, and density.—Accurate determinations of the optical properties of the biotites in the paragneiss are very difficult and in many specimens practically impossible. Most specimens are deeply colored—dark greenish-brown, red, or black—verging on opaque except in the thinnest flakes. All exhibit very weak dispersion and strong absorption; hence indices of refraction are hard to obtain and liable to considerable error ($\pm.004$). These characteristics of the biotites are not nearly so apparent in thin section as in immersion oils. The difficulty in obtaining systematic consistent optical constants in these specimens is surprising in view of their fairly uniform chemical characteristic and their color in thin section and in purified samples.

Because of the desirability of recording optical data, 7 of the biotites which seemed most amenable to study were submitted to Jewell Glass of the U. S. Geological Survey. She reports the same difficulties we observed and was unable to obtain precise data. Consequently, the optical data recorded in Table 13 are confined to N_z and color.

Density determinations were made on 10 biotites—3 of these are from the Emeryville area (Group 1) and 3 from Colton (Group 5). Determinatious were made with a micropycnometer on samples ranging from 2.7 to 7.2 grams. Each value listed is the mean of 2 determinations, whose mean deviations ranged from .003 to .023 (Table 13).

X-ray procedures.—Some of the data on crystal structure of the biotites were contributed to this study by A. A. Levinson, B. Raychaudhuri, and H. S. Yoder, Jr. Yoder kindly undertook powder-diffraction studies of 10 biotites from the paragneiss in an effort to distinguish the polymorphs present. Regarding preparation of the samples and interpretation, Yoder (written communication, 1954) states:

> "Portions of the samples were ground in acetone for two hours and the −325 fraction was x-rayed in a cell mount in order to reduce the preferred orientation. Unfortunately, the preferred orientation was such that the reflections in the critical region of 2 θ necessary to distinguish the polymorphs were not as sharp as desired. For this reason the identification of the polymorphs is not as certain as I would like to make from an X-ray powder pattern".

Subsequent to the powder-diffraction studies, Levinson offered to undertake single-crystal studies on the coarsest-grained sample of biotite separated from the paragneiss. Specimen Qb 228 was selected for it contains flakes up to 0.3 mm in diameter. Levinson definitely was able to identify the single flake of Qb 228 that he studied as a 2 layer monoclinic (2 M) polymorph with moderate diffuse scatterings. Levinson also studied the powder pattern of Qb 228, using filtered Fe radiation and an exposure of about 12 hours in a camera of 114.59 mm diameter. He notes:

> "The pattern was excellent and showed some very weak lines in the range d = 4.5 Å–2.65 Å, which might be characteristic of the 2 M structure, if the one flake from which the 2 M structure was obtained is characteristic of the rest of the specimen."

Studies of cell dimensions in 8 biotites were made with the assistance of B. Raychaudhuri. The Norelco Geiger-Counter X-ray spectrometer was used with copper radiation. X-ray mounts of micas were prepared by the method described by Yoder and Eugster (1954, p. 162). Sodium chloride was used as an internal standard.

All mica mounts were studied for (004) and (005) reflections. Each was exposed and scanned 6 times to determine reproducibility of measurements. Under these conditions the maximum deviation recorded is .0022 for c sin B (004) or (005) and .0005 for d in Å. Larger errors are introduced if the same sample is emptied and re-

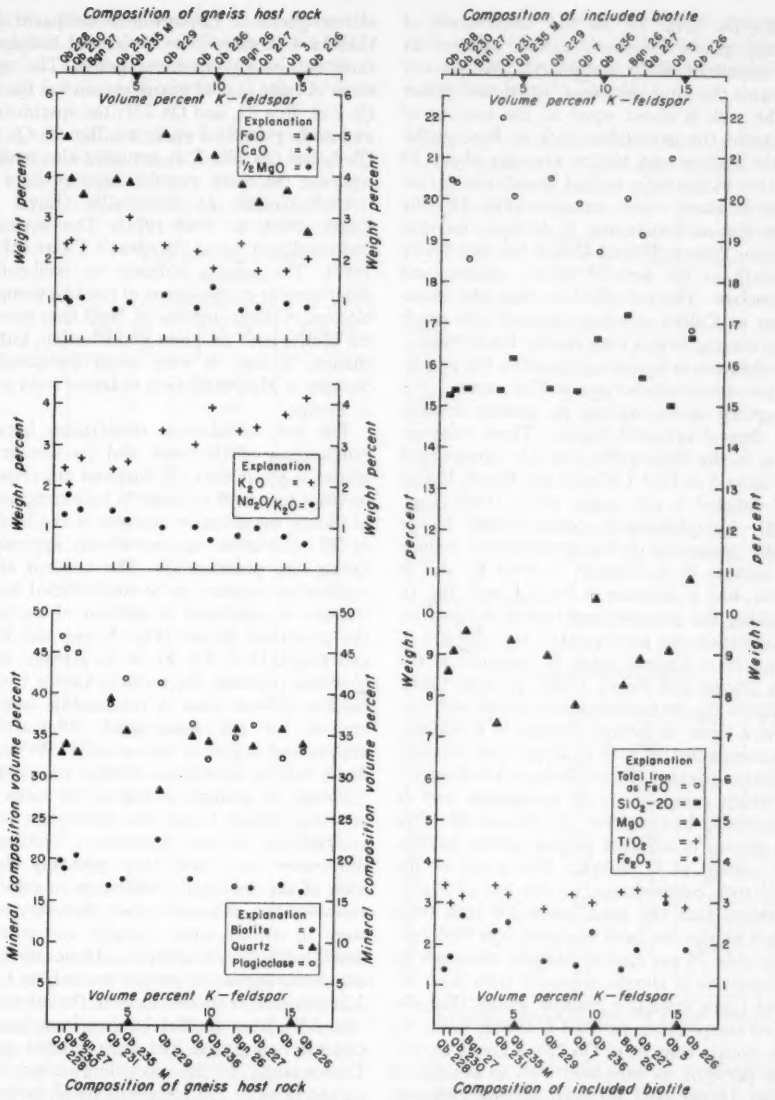

FIGURE 5.—VARIATIONS IN THE MINERALOGICAL COMPOSITION OF THE PARAGNEISS FROM THE EMERYVILLE AREA AND IN THE COMPOSITION OF CONSTITUENT BIOTITES, PLOTTED AGAINST INCREASING K-FELDSPAR CONTENT

packed, but these do not exceed ±0.008 for *c* sin B (004) or (005) and ±0.002 for *d* in A.

Occurrence

Biotite is ubiquitous in the paragneiss complex, and its concentration in the rock specimen varies with degree of metamorphism and amount of granitization. These relationships have been discussed in Part I of this study and are briefly summarized in Figures 1 and 5 of this paper. Biotite averages about 18 per cent by volume in the least altered paragneiss at

Emeryville (Fig. 1). As the temperature of metamorphism increases, garnet appears at the expense of biotite and quartz. Northeast of Edwards the total volume of biotite and garnet in the rock is about equal to the amount of biotite in the garnet-free rock at Emeryville. In the Colton area biotite averages about 10 per cent of the rock, so that biotite and garnet occur in about equal amounts (Fig. 1). The grain size of biotite and K feldspar increase between Emeryville and Colton but not nearly so much as the sizes of garnet, quartz, and plagioclase. The net effect is that the paragneiss at Colton is coarser-grained and much more inequigranular than that at Emeryville.

A decrease in biotite accompanies the granitization of least altered gneiss. The parts of the paragneiss most replaced by granite contain less than 3 per cent biotite. These relationships in the Emeryville area are summarized in Figure 3 of Part I (Engel and Engel, 1958a) and detailed in our earlier study (1953, Figs. 5, 6). Complementary compositional trends in the paragneiss during granitization include an increase in K feldspar, in total K, and in K/Na, and a decrease in Fe, Ca, and Mg. In graphing the compositional trends induced by granitization we have plotted the changes in composition against total K feldspar in the rock (Engel and Engel, 1958a, p. 1403–1411).

Where the granitization is incipient and uniform, a three- to fivefold increase in K feldspar is accompanied by a 15 to 20 per cent decrease in biotite content. This is indicated in Figure 5 in which compositions of paragneiss and of constituent biotite from the Emeryville area are plotted in adjacent graphs against increasing content of K feldspar. The graph of the total rock compositions, on the left of Figure 5, shows that the most granitized rock from which biotite has been analyzed (Qb 226) contains only 15 per cent K feldspar, although its composition is clearly displaced from least altered types toward a granitic gneiss. (For detailed composition, *see* Part I, 1958a, Table 9.) The biotite content of Qb 226 is about 16 volume per cent, as compared with an average of about 18 per cent for least altered gneiss in the Emeryville area (Fig. 1).

Considering the systematic changes in total rock composition induced during incipient granitization at Emeryville, it is noteworthy that no clearly defined relationships exist between composition of gneiss and constituent biotite in any area or group. This is suggested from inspection of the right side of Figure 5 and by analyses of variance in which the amount of each element in 4 biotites in least altered gneiss at Emeryville is compared with that in 4 biotites from associated incipiently (and uniformly) granitized gneiss. The specimens of least altered gneiss chosen are Qb 228, Qb 230, Bgn 27, and Qb 231; the specimens of uniformly granitized gneiss are Bgn 26, Qb 227, Qb 3, and Qb 226. This grouping also tends to separate the more greenish biotites from the reddish biotites as Emeryville (Engel and Engel, 1953, p. 1068–1071). The statistical test employed was a "Student's T-test (Mood, 1950). The results indicate no well-defined differences in compositions of the two groups of biotites. A slight increase of MgO may occur in the biotite with incipient granitization, but the change, if real, is very small compared to changes in MgO in biotites as traced from group to group.

The lack of obvious correlations between composition of the rock and its constituent biotite suggests that (1) incipient granitization involves too small a change in bulk composition to induce recognizable changes in the biotites, or (2) equilibrium was not closely approached during the granitization. The concept of inequilibrium appears to be contradicted by the changes in color and in amount of biotite in the granitized gneiss (Fig. 5; *see also* Engel and Engel, 1953, Fig. 8). In an attempt to explore the problem, the rocks in Group 1 whose biotites differed most in composition were examined for any meaningful differences in texture and degree of homogeneity. We cannot find a definite correlation between composition, lithology, or geologic setting of the rocks constituting Group 1 and real differences in the composition of the constituent biotites. If differences exist—and they probably do, in view of the systematic differences in color and concentration of biotite—their discovery would seem to require larger samples and more refined methods of analysis. These preceding comments appear to pertain not only in Group 1 but in each of specimens from the other areas where we have studied biotites from both incipiently granitized and least altered gneiss. Consequently, in the succeeding section where comparisons of the compositions of biotites of different groups are made, the biotites from least altered gneiss and incipiently granitized gneiss of any one group are treated as indistinguishable in composition.

Composition Related to Metamorphic Grade

Between Emeryville and Colton the biotites in the paragneiss increase in TiO_2, MgO, F, Ba, Cr, Cu, Ni, V, Zr and decrease in Fe_2O_3, FeO,

total Fe, Mn, H_2O, and Sc. These changes in composition are shown in Tables 13–15 and in Figure 6. The mean compositions of biotites in Groups 1, 2, 3, and 5 appear in Table 15.[8] Data from Group 4 are omitted in Table 15 inasmuch as only 1 biotite was analyzed *in toto* (Table 13). The number of biotites constituting the means varies as follows: 12 in Group 1 (Emeryville area), 3 in Group 2, 4 in Group 3, and from 3 to 6 in Group 5[9] (Colton area).

TABLE 14.—TRACE ELEMENTS, TiO_2, FeO, COLOR, AND DENSITY FOR THREE ADDITIONAL BIOTITES, COLTON AREA (GROUP 5), NEW YORK
Trace elements in parts per million

Element	Colton Area Group 5 Specimens		
	Bgn 20	Bgn 23	Bgn 24
Ba	1800	1800	2000
Co	50	40	50
Cr	350	250	200
Cu	40	30	100
Ga	50	60	40
Mn	80	100	100
Nb	20	50	40
Ni	120	80	100
Sc	20	15	30
Sr	20	30	30
V	750	400	400
Y	10	30	30
Yb	4	4	2
Zr	200	200	300
Oxide (weight %)			
TiO_2	5.01	4.87	4.93
FeO	15.03	14.38	13.93
Color	drbr	drbr	drbr
Density	3.148	3.153	3.151

The atomic ratios of elements in each specimen are listed in Table 16. Differences between the mean atomic ratios in the biotites of Groups 1 and 5 are listed in Table 17.

TABLE 15.—AVERAGE COMPOSITION OF BIOTITES FROM EMERYVILLE, EDWARDS, RUSSELL, AND COLTON AREAS, NEW YORK*

	Emeryville area (Group 1)	Edwards area (Group 2)	Russell area (Group 3)	Colton area (Group 5)
SiO_2	35.80	35.72	36.95	36.93
TiO_2	3.15	4.21	4.39	5.22
Al_2O_3	18.64	17.25	17.59	17.38
Fe_2O_3	2.02	1.67	.97	1.30
FeO	17.98	19.02	17.72	14.30
MnO	.20	.09	.00	.00
MgO	9.01	8.97	9.48	11.81
CaO	.12	.29	.43	.61
Na_2O	.29	.16	.46	.29
K_2O	9.09	8.99	8.21	8.54
H_2O^+	3.40	3.24	3.56	3.25
F	.31	.35	.33	.45
Number of specimens	12	3	4	3
Trace Elements (ppm)				
Ba	580	717	888	1766
Co	40	40	38	49
Cr	145	153	173	275
Cu	27	77	45	54
Ga	48	51	53	57
Mn	1430	630	185	87
Nb	34	33	35	35
Ni	77	73	88	107
Sc	68	54	47	20
Sr	17	17	16	23
V	328	393	439	558
Zr	77			190

* For individual analyses, *see* Tables 13 and 14.

Efforts to relate our data to available data on metamorphic biotites from pelitic schists in other terranes emphasize the very fragmentary nature of knowledge of common metamorphic rocks and minerals as well as the hazards of direct comparisons of published analyses of minerals. Most studies that adequately specify the lithology and geologic setting of pelitic schists say little or nothing about the chemical composition of the rocks or the constituent minerals. Conversely most pub-

[8] The differences in composition of biotites between Groups 1 and 5 are verified by statistical analyses, using the Student's t-test (Mood, 1950, Chapter 9). The results show that for TiO_2, FeO, MgO, MnO, Ba, Cr, Cu, Ni, Zr, and Sc the probability of the difference of the means between Group 1 and Group 5 being due to chance are less than 0.01. For F and H_2O the probability is about or slightly less than 0.05.

[9] The 3 specimens from Table 14 are included in the means for all trace elements.

lished analyses of one or more minerals do not give either a mode or a chemical analysis of the host rock, much less details of its geologic occurrence. Hence it is difficult to place most of the metamorphic minerals cited in the literature in any sequence of increasing temperature of formation, or to relate compositions of constituent minerals to the composition or geologic occurrence of the parent rock.

The lack of data on analytical error in published analyses of minerals is an equally serious impediment to comparison of compositions. Most estimates of purity of the sample made by microscopic inspection are far too optimistic. Petrographic studies to determine purity of biotites, for example, cannot always establish whether several per cent of chlorite, or of other sheet-structure silicates, exists in interlayer positions. X-ray analyses will indicate several per cent or more of chlorite, but these are rarely made on mineral concentrates submitted for analysis. The problem of how to distinguish less than 3 or 4 per cent interlayer chlorite would seem to require new and special techniques. Until mineral concentrates are more carefully studied, conclusions regarding their purity will probably remain articles of faith and hope. Despite these objections crude com-

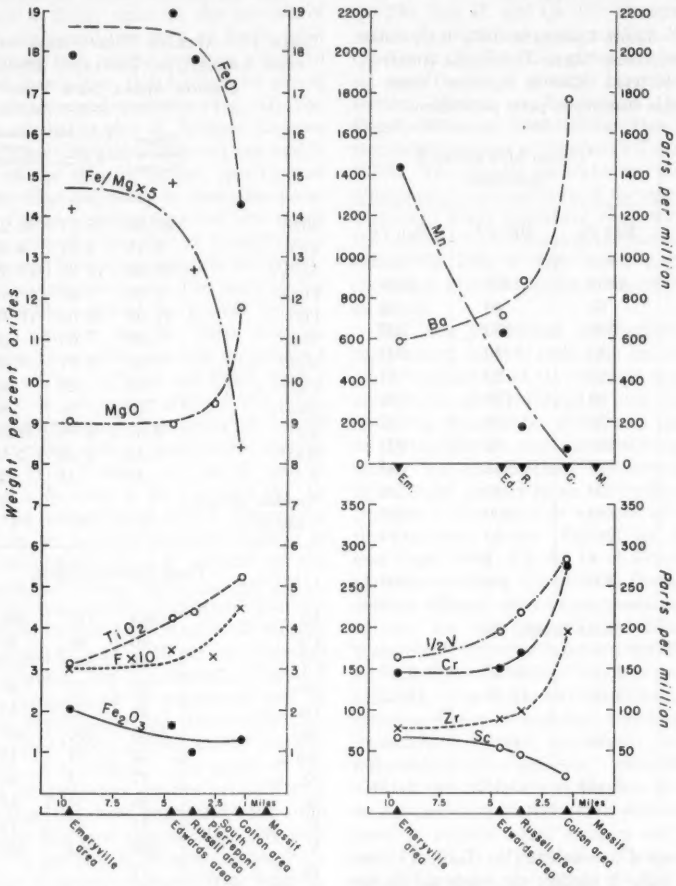

FIGURE 6.—VARIATIONS IN COMPOSITION OF BIOTITES FROM THE PARAGNEISS PLOTTED AS A FUNCTION OF INCREASING GRADE OF METAMORPHISM BETWEEN EMERYVILLE AND COLTON, NEW YORK

TABLE 16.—COMPOSITION OF BIOTITES IN THE ADIRONDACK PARAGNEISS, EXPRESSED IN ATOMIC RATIOS*

TABLE 16.—COMPOSITION OF BIOTITES IN THE ADIRONDACK PARAGNEISS, EXPRESSED IN ATOMIC RATIOS*

Atomic ratios	West Balmat (Group 1A)		Emeryville (Group 1)										Edwards (Group 2)			Russell (Group 3)				South Pierrepont (Group 4)	Colton (Group 5)		
	Bgn 26	Bgn 27	Qb236	Qb 235M	Qb 3	Qb 7	Qb 230	Qb 229	Qb 231	Qb 226	Qb 227	Qb 228	Qb 233	Qb 100	Qb 216	Bgn 6	Bgn 9	Bgn 14	Bgn 4	Bgn 25	Bgn 18	Bgn 21	Bgn 19
Tetrahedral																							
Si	2.82	2.68	2.78	2.76	2.66	2.68	2.69	2.69	2.73	2.87	2.69	2.69	2.72	2.76	2.71	2.74	2.79	2.80	2.75	2.65	2.82	2.71	2.71
Al	1.18	1.32	1.22	1.24	1.34	1.32	1.31	1.31	1.27	1.13	1.31	1.31	1.28	1.24	1.29	1.26	1.21	1.20	1.25	1.35	1.18	1.29	1.29
	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00
Octahedral																							
Al	.45	.37	.34	.32	.25	.49	.39	.37	.39	.62	.44	.37	.36	.20	.29	.29	.44	.31	.25	.24	.35	.23	.23
Fe^{+++}	.08	.15	.13	.15	.17	.09	.10	.09	.14	.11	.11	.08	.14	.07	.08	.07	.04	.05	.06	.09	.09	.06	.06
Fe^{++}	1.19	1.02	1.05	1.13	1.17	1.17	1.20	1.21	1.28	.99	1.10	1.23	1.03	1.40	1.23	1.11	.97	1.19	1.17	1.03	.85	.96	.87
Mg	.93	1.08	1.17	1.06	1.03	.88	1.03	1.01	.84	.72	1.00	.98	1.08	.95	1.04	1.09	1.08	.97	1.08	1.35	1.24	1.25	1.45
Mn	.01	.02	.02	.02	.02	.01	.02	.01	.02	.02	.02	.01	.00	.02	.02	..	..	..	..	..	..	..	..
Ti	.19	.18	.17	.18	.18	.18	.17	.16	.20	.17	.19	.20	.23	.24	.25	.26	.22	.27	.24	.22	.28	.31	.29
	2.85	2.82	2.88	2.86	2.82	2.82	2.91	2.85	2.87	2.63	2.86	2.87	2.84	2.88	2.89	2.82	2.75	2.79	2.80	2.93	2.81	2.81	2.90
Large cation																							
Ca	.01	.02	..	.01	.01	..	..	..	..	.03	.02	.01	.03	.03	.02	.02	.07	.03	.02	.01	.08	.06	.02
Na	.03	.06	.01	.02	.05	.06	.05	.03	.08	.06	.06	.01	.03	.03	.01	.05	.08	.07	.05	.06	.05	.06	.01
K	.90	.89	.93	.93	.84	.90	.86	.90	.90	.88	.78	.91	.85	.89	.89	.80	.78	.80	.76	.87	.78	.83	.82
	.94	.97	.94	.96	.90	.96	.91	.93	.98	.97	.86	.93	.91	.95	.92	.87	.93	.90	.83	.94	.91	.95	.85
OH	1.60	1.75	1.59	1.68	1.99	1.74	1.72	1.78	1.59	1.77	1.75	1.75	1.64	1.64	1.67	1.75	1.75	1.68	1.96	1.71	1.53	1.64	1.67
F	.06	.09	.08	.07	.05	.06	.09	.10	.11	.07	.06	.08	.08	.12	.06	.09	.08	.06	.07	.09	.09	.10	.12
	1.66	1.84	1.67	1.75	2.04	1.80	1.81	1.88	1.70	1.84	1.81	1.83	1.72	1.76	1.73	1.84	1.83	1.74	2.03	1.80	1.62	1.74	1.79

* For method of calculation of atomic ratios *see* Stevens (1946)

parisons are attempted with available data on metamorphic biotites from pelitic schists in other terranes. The data presented by Barth (1936), DeVore (1955a; 1955b), and Phillips (1930) are especially pertinent. All the difficulties of comparisons noted above exist here. The most ambitious attempt to relate composition of biotites to grade of metamorphism is the work of DeVore. Unfortunately, the analyses of major as well as minor and trace elements posted by DeVore (1955a) were done by emission spectrography. Totals of analyses of major and minor constituents of biotites vary from 100 per cent by as much as 5 per cent. The analytical precision for at least some elements is, therefore, quite low. DeVore generously loaned us powders of these biotites so that we might attempt to relate some of his analytical data to our own. Impurities in his samples range from 3 to 6 per cent. None of the samples was large enough to permit gravimetric analyses for major and minor elements. Spectrographic analyses for trace elements were made on these biotites along with some of our "standard" biotites. The differences between the values DeVore reports for his samples and the numbers we obtain are large enough to obscure many compositional trends. A second complication is that most of DeVore's biotites are from garnet-free hornblende- and pyroxene-bearing metamorphic rocks differing appreciably from the paragneiss in bulk composition and in the coexisting mafic phase. Consequently, the trends in the composition of the biotites studied by DeVore, especially in Fe, Mg, and chemically coherent trace elements, cannot be expected to be similar to those we observed in the paragneiss.

In spite of the difficulties and shortcomings cited above we have attempted a comparison of DeVore's data with our own. In Table 18 are posted the average, maximum, and minimum values of each major oxide and minor element for two groups of biotite specimens listed by DeVore (1955a, Table 5) as from respectively the "amphibolite facies" and "granulite facies." Alongside these data are average, maxima, and minima for these elements from biotites in Groups 1 and 5 of the Adirondack paragneiss. This arrangement assumes that the biotites of Group 1 represent about the same physical and chemical conditions of metamorphism as do the biotites of DeVore's "amphibolite facies." The biotites of our Group 5 are inferred to correspond roughly, in environment of origin, to the biotites from rocks referred to by DeVore as "granulite facies." Some of the paired means from the two sets of data are surprisingly close, but because of the probable errors inherent in this comparison the results must be treated cautiously.

Detailed comparisons of the composition of biotites in the paragneiss with those from igneous and metamorphic rocks seems unjustifiable, but a gross comparison is attempted in Figure 7 (after Heinrich, 1946). Heinrich used a ternary diagram to plot the range in composition of biotites from pegmatites, granites, gneisses and schists, diorites, gabbros, basalts, and ultramafics. In general the more

TABLE 17.—MEAN COMPOSITIONS OF BIOTITES IN GROUPS 1 AND 5 EMERYVILLE AND COLTON AREAS, NEW YORK, EXPRESSED IN ATOMIC RATIOS

Area and group	Value	Tetrahedral			Octahedral							Large cation				Hydroxyl halogens		
		Si	Al	Sum	Al	Fe^{+++}	Fe^{++}	Mg	Mn	Ti	Sum	Na	K	Ca	Sum	OH	F	Sum
Emeryville* (Group 1)	x̄	2.72	1.28	4.00	.40	.12	1.14	.98	.02	.18	2.84	.04	.88	.01	.94	1.74	.08	1.81
Colton† (Group 5)	x̄	2.75	1.26	4.00	.27	.07	.89	1.31	tr	.29	2.84	.04	.81	.05	.90	1.61	.10	1.72

tr = trace (<.01)
* Eleven specimens
† Three specimens

TABLE 18.—COMPARISONS OF COMPOSITIONS OF BIOTITES IN THE PARAGNEISS WITH COMPOSITIONS OF METAMORPHIC BIOTITES DISCUSSED BY DEVORE (1955a)

Oxide or element	Mean Number of samples in parentheses		Analyst
	Amphibolite facies	Granulite facies	
SiO_2	35.80 (12)	36.93 (3)	C. Engel
	37.50 (4)	36.33 (12)	Joensuu*
TiO_2	3.15 (12)	5.08 (6)	C. Engel
	2.48 (4)	2.08 (12)	Joensuu
Al_2O_3	18.64 (12)	17.38 (3)	C. Engel
	18.50 (4)	16.13 (12)	Joensuu
Total Fe	15.25 (12)	12.03 (3)	C. Engel
	13.55 (4)	17.41 (11)†	Joensuu
MgO	9.01 (12)	11.81 (3)	C. Engel
	10.85 (4)	11.03 (12)	Joensuu
H_2O+	3.40 (12)	3.25 (3)	C. Engel
	(In parts per million)		
Co	40 (12)	49 (6)	Chodos and Godijn
	75 (4)	47 (12)	Joensuu
Cr	145 (12)	275 (6)	Chodos and Godijn
	214 (4)	62 (12)	Joensuu
Cu	27 (12)	54 (6)	Chodos and Godijn
	52 (4)	30 (12)	Joensuu
Mn	1430 (12)	87 (6)	Chodos and Godijn
	1250 (4)	2013 (12)	Joensuu
Ni	77 (12)	107 (6)	Chodos and Godijn
	381 (4)	50 (12)	Joensuu
Sc	68 (12)	20 (6)	Chodos and Godijn
	29 (4)	15 (12)	Joensuu
Sr	17 (12)	23 (6)	Chodos and Godijn
	87 (4)	80 (12)	Joensuu
V	328 (12)	558 (6)	Chodos and Godijn
	369 (4)	203 (12)	Joensuu
Zr	77 (12)	190 (6)	Chodos and Godijn
	55 (4)	70 (12)	Joensuu
$\frac{Mg}{Mg + Fe}$	0.26 (12)	0.37 (3)	C. Engel
	0.33 (4)	0.26 (11)†	Joensuu

* In DeVore (1955a). The means are probably low in that "trace" of any element in a given sample is calculated as zero.

† Sample M6 omitted because of probable misprint

mafic the parent rock and the higher its temperature of formation, the more enriched the constituent biotite in MgO relative to FeO + MnO, and Fe_2O_3. We have added the biotites from the paragneiss to the ternary plot. They span the fields between (1) biotites from granites and those from diorites and (2) biotites in gneisses and schists. As might be predicted, biotites from the more "felsic" paragneiss at Emeryville are more like biotites from granites; and the biotites from the basified paragneiss at Colton are much like biotites found in diorites.

Silica.—The SiO_2 content of the biotites appears to increase slightly between biotites of Group 1 and Group 5, but it is not highly

sensitive to the recorded differences in metamorphism. A similar insensitivity of Si to temperature has been observed in biotites in igneous rocks (Nockolds, 1947).

morphic grade from 4.82 per cent TiO_2 in the "low-grade" biotite to 2.48 per cent in "high-grade" biotite (Barth, 1936, p. 783). Barth does not record the amounts of TiO_2 in the

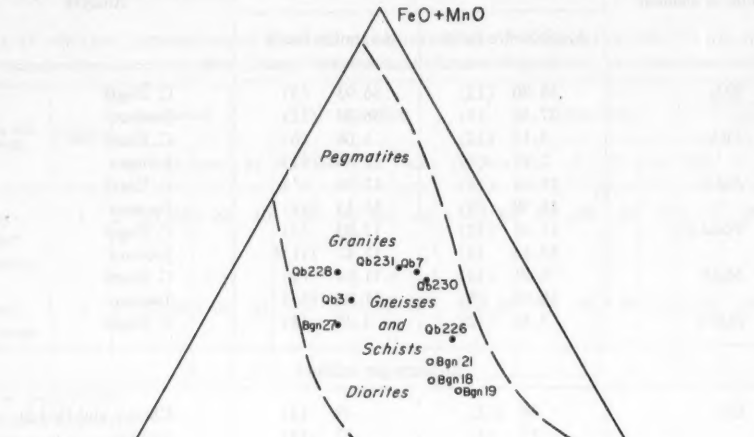

FIGURE 7.—COMPARISON OF THE COMPOSITIONS OF BIOTITES IN THE PARAGNEISS WITH COMPOSITIONS OF BIOTITES FROM IGNEOUS AND METAMORPHIC ROCKS (after Heinrich, 1946)

Dashed lines indicate limits of fields of composition of biotites. Solid dots are biotites from Emeryville (Group 1). Open circles are biotites from Colton (Group 5).

Titanium.—TiO_2 content increases in the biotite of the paragneiss with increasing temperature of metamorphism. This probably is true in many metamorphic terranes. This conclusion follows from the relations of TiO_2 content to colors of metamorphic biotites (as discussed subsequently) although an increase of Ti with increasing grade of metamorphism is not indicated by all the published analyses. Analyses of "low"- and "high"- grade biotites from the Green Beds of the Scottish Dalradian (Phillips, 1930) show an increase in TiO_2 from 0.23 to 1.72 (weight per cent). Conversely, two analyses of biotite from Dutchess County show a decrease in TiO_2 with increasing metaparent rock or the identity of associated minerals.

Ramberg (1952, p. 75) and DeVore (1955a) have stated that an increase in TiO_2 in biotites is the rule with increasing temperature of metamorphism, but the analytical data DeVore posts do not confirm his conclusion. This may be seen from inspection of Table 18. The mean of TiO_2 in DeVore's biotites from the "amphibolite facies" is 2.48 weight per cent, whereas the mean for biotites in his "granulite facies" is 2.08 weight per cent (DeVore, 1955a, p. 167).

The relatively high concentrations of Ti in the biotites of the Adirondack paragneiss as

well as the increase in TiO_2 from Groups 1 to 5 are a reflection of the fact that the Adirondacks are a titanium-rich province. Probably the amounts of Ti recorded approach the maximum saturation amounts for this type of system at the physical conditions pertaining, inasmuch as ilmenite ($FeTiO_3$) invariably is present as scattered grains in the host rock.

The marked increase of TiO_2 in the biotites of Groups 3, 4, and 5 is not accompanied by a decrease in Si. This fact seems to argue against diadochic replacement of Si^{+4} by Ti^{+4}, as suggested by Kunitz (1936) and many others. Presumably the bulk of the Ti in the Adirondack biotites is trivalent and replaces aluminum and ferric iron.

Aluminum.—The total Al content in biotites in the paragneiss may decrease with increasing rank of metamorphism. Inasmuch as the biotites are rich in Al this change appears in the octahedrally co-ordinated Al (Table 16). There are, however, large variations in Al content solely within the biotites of the Emeryville area where specimens Qb 236, Qb 235M, and Qb 3 contain essentially the same amounts of Al as do biotites at Colton. One of the Edwards biotites (Qb 100) also seems to be no more aluminous than those at Colton.

There is no highly systematic relation between the variations in Al and those of Si or Ti although the least aluminous biotites at Emeryville are richer in Fe^{+++} than the average.

Iron.—The total Fe content and amounts of Fe^{+++} and Fe^{++} in the biotites decrease markedly with increasing temperature of metamorphism. The mean values of Fe_2O_3 and FeO in the biotites near Emeryville (Group 1) are, respectively, 2.00 and 18.00 weight per cent; in the Colton area (Group 5) the means are 1.3 and 14.3 (Table 15). The ratio of Fe_2O_3 to FeO also decreases (0.11 to 0.09) with increasing grade of metamorphism.

The depletion of Fe in the biotites with increasing temperature of metamorphism is opposite the trends suggested by DeVore (1955a, p. 163–164) from calculations of analyses of both metamorphic and igneous biotites. One possible reason for the differences in trends of Fe may lie in the mineral assemblages in which the biotites are found. In the Adirondack assemblage, garnet coexists with the biotite in the higher grade of metamorphic groups. The decrease in total iron in the biotite reflects the partitioning of iron—and magnesia—between the biotite and coexisting garnet.[10] This partitioning effect is discussed further after the data on garnets are presented. It may be significant that the biotite with the highest Mg/Fe ratio shown in DeVore's analyses is from one of the rare rocks in which garnet coexists with biotite (DeVore, 1955a, p. 166, specimen 370–46).

In the analyses of "low" and "high-grade" metamorphic biotites posted by Barth, both total Fe and FeO decrease with increasing grade of metamorphism, whereas Fe_2O_3 appears to increase slightly (Barth, 1936, p. 783. specimens 1 and 2). Garnet coexists with biotite in the specimen that Barth described as "high grade."

Garnet also seems to occur in the rock from which the "low" and "high-grade" biotites were separated by Phillips (1930). In these two biotites, Fe_2O_3 and total Fe appear to decrease with increasing grade, but FeO increases. Possibly part or most of these reported differences are due to the kinds of errors of measurement we reported in the discussions of FeO determinations.

Magnesium.—A marked increase in Mg occurs in biotite during metamorphism at higher temperatures. Consequently, in the biotites an antipathetic relation exists between Mg and Fe^{++} as a function of increasing temperature of metamorphism (Fig. 6). The FeO:MgO ratio of the mean biotite decreases from about 2.0 at Emeryville to 1.2 at Colton. The FeO:MgO ratio of the low- and high-grade biotites analyzed by Barth (1936, p. 783) changes similarly from 2.1 to 1.5. Phillips' analyses also show a decrease in FeO:MgO of 1.1 to 0.9 with increasing rank of metamorphism. In all these examples, the higher grade of biotite coexists with garnet.

Manganese.—Biotites in the Colton area (Group 5) contain only about $\frac{1}{15}$ the Mn present in the biotites of the Emeryville area. This decrease from more than 1300 to 88 parts per million Mn, with increasing grade of metamorphism, is correlative with the decrease in Fe^{++} in the biotites from 14.0 to 11.03. Presumably the Mn:Fe^{++} ratio decreases from .01 in Group 1 to .0008 in Group 5. Goldschmidt (1954, p. 627–631) has noted that in igneous rocks the ratio of Mn to Fe^{++} similarly decreases with increasing basicity of the rocks and with increasing temperature of formation.

[10] Magnetite occurs as an accessory, but it constitutes less than 0.1 of the rock.

There are few data on the amount of Mn in metamorphic biotites in paragneisses of other regions. DeVore (1953, p. 1412) has stated that the Mn content of biotites increases with increasing temperature of formation. His published figures tend to support this conclusion (paper, Table 18), although variations in Mn content of individual specimens are very large (DeVore, 1955a, p. 166).

The Mn content of the "high-grade" biotite described by Barth is not reported, but 0.5 MnO is given for the sample from a "low-grade" schist. In the two analyses of metamorphic biotites that Phillips published, the Mn content decreases with increasing rank of metamorphism from 0.53 MnO in "low-grade" biotites to "nil" MnO in "high-grade" biotite. If "nil" MnO is on the order of 0.05 weight per cent, this represents a decrease in Mn analogous to that in the biotites of the paragneiss.

Calcium.—One-tenth to several tenths of a per cent CaO recorded in the biotite analyses must be largely intrinsic. No strong trend in CaO concentration is apparent, although 2 out of 3 of the Colton biotites are higher, by a factor of 3, than 90 per cent of all other specimens.

Potassium and sodium.—The K and Na contents posted for the biotites (Tables 13, 16) are slightly low because of 0.5 to 1.5 per cent chlorite impurity, as noted in the discussion of Relation of Biotite Composition to Crystal Structure. When allowance is made for the chlorite, however, no marked trend is apparent in the Na and K content or in Na/K with changes of metamorphic grade.

Fluorine.—The fluorine content of the biotites in the gneiss increases 50 per cent with increasing temperature of metamorphism (Group 1 to Group 5). This increase in F is not accompanied by a clearly defined change in the OH content, although OH probably decreases slightly as F increases. These changes in F with increasing grade of metamorphism are directly opposite those noted by DeVore in his recalculations (1955a, Fig. 2; Table 3) of Nockolds' data (1947) on igneous biotites. Fluorine has been determined on only a few biotites from intermediate to high-grade metamorphic rocks (*see*, for example, Barth, 1936, p. 783), hence there is little basis for comparison. The biotites from the paragneiss are low in fluorine when compared to some metamorphic biotites (Lee, 1958).

Barium.—The Ba content of biotites in the gneiss triples between Group 1 and 5, with increasing grade of metamorphism. According to classical theory the Ba replaces K diadochically in the mica structure.

There are few published analyses of Ba in metamorphic biotites with which our analyses may be compared. Data on igneous biotites are of interest in that the Ba content appears to increase with increasing temperature of growth of the biotite (Von Engelhardt, 1936). A maximum of about 3000 parts per million Ba in igneous biotites is reported by Von Engelhardt—almost twice the Ba content of the biotites in the Colton area.

Cobalt and nickel.—Biotites in Group 1 average 40 and 77 parts per million Co and Ni respectively. Both the Co and Ni content of the biotites may increase, with increasing grade of metamorphism, to about 50 and 107 parts per million in the Colton area. Consequently, the Co:Ni ratio remains about 0.5 in biotites from the several groups; moreover, the Co:Mg and Ni:Mg ratios also remain about constant, at .0007 and .0015 (Fig. 6). The ratios of $Co:Fe^{++}$ and $Ni:Fe^{++}$ vary appreciably, increasing markedly with increase of metamorphic grade. The ratio of Co:Ni of 0.5 in the biotites is about that reported from many igneous rocks (Rankama and Sahama, 1950, p. 682; Nockolds and Mitchell, 1948, p. 541) and from the hydrolyzate sediments (Rankama and Sahama, 1950). There is no doubt of a more consistent relation between Co and Mg, and between Ni and Mg, than between Co, Ni, and Fe^{++}. This also seems to be true in igneous rocks (Sandell and Goldich, 1943). Co and Ni values of 75 and 381 parts per million are reported by DeVore (1955a) from biotites of the amphibolite facies, and 47 and 50 parts per million Co and Ni from biotites of the granulite facies; these decreases in Co and Ni with increased metamorphic grade appear, however, to be accompanied by increases in both Fe and Mg (Table 18).

Chromium.—The chromium content of the biotites appears to double between Groups 1 and 5 (Tables 13, 15). Although this increase parallels that of Mg, it is far more pronounced, and it is inversely related to the trends in concentration of Fe^{+++} and Fe^{++} (and Al?) in the biotites. The only other major or minor element in the biotites that is analogous to Cr in rate of increase with increasing grade of metamorphism is Ti (Table 15). According to Goldschmidt's concepts of "camouflage," "capture," and "admittance" (1954), Cr^{++} may substitute for Ti^{+++} where the deficient positive charge of Cr could be balanced by other substitutions.

Scandium.—The Sc content of the biotites

decreases from a mean of 68 in Group 1 (Emeryville) to 20 in biotites near Colton. This decrease in Sc with increasing grade of metamorphism is opposite the trend of Mg but may be related in part to the parallel decrease in Fe^{++}.

Color

The color changes in biotites from various types and areas of paragneiss are striking and correlative in a general way with major changes

TABLE 19.—RELATION OF COLOR TO TiO_2, TOTAL Fe (AS FeO), AND MgO IN BIOTITES FROM THE PARAGNEISS

Group	TiO_2	Total Fe (as FeO)	MgO	$\frac{TiO_2}{FeO}$	$\frac{TiO_2:MgO}{FeO} \times 10$	Color
1 (500° C)	3.16*	19.80	9.01	.16	.18	Greenish brown
2	4.21	20.52	8.97	.21	.23	Brown
3	4.51	18.59	9.48	.24	.25	Reddish brown
4	4.73	17.78	10.12	.27	.26	Reddish brown
5 (600° C)	5.04†	16.06	11.86	.31	.27	Deep reddish brown

* Specimens in Groups 1 and 1A, Table 13
† Includes Bgn 24 and Bgn 20 (Table 14)

Actually, the $Sc:Fe^{++}$ ratio decreases from .0005 in Group 1 to .0002 in Group 5. A similar decrease in Sc content of biotites with increasing grade of metamorphism is reported by DeVore (1955a, p. 166) although the Fe^{++} and Mg values he posts increase with metamorphic rank (Table 18). Oftedal (1943) has noted that there is also a decrease in Sc in biotites with increasing temperature of origin. He records from 10 to 30 parts per million Sc in biotites from granitic igneous rocks, about 20 parts per million in biotites of "regionally metamorphosed rocks", and only 7 to 10 parts per million in basic igneous rocks. On the basis of these figures Oftedal concluded that the Sc content of biotites could be used as a geologic thermometer—the higher the Sc content the lower the temperature of formation. (*See also* Ingerson, 1955, p. 364.) Our data certainly sustain this generalization.

Vanadium.—The V content of the biotites appears to increase from a mean of 328 parts per million in Group 1 to 558 in Group 5. This seeming change in concentration of V by a factor of nearly 2 is opposite to the recorded trends in concentrations of Al and Fe^{+++} but is sympathetic with the increase in Ti and with temperature of crystal growth.

Zirconium.—The posted mean values of Zr in the biotites of each group (Tables 13, 15) are probably intrinsic, and not due to adulteration by microcrystalline zircon. Assuming the concentrations of Zr observed are correct at 77 to 190 parts per million (Table 15) these are essentially the same Zr values reported by DeVore for biotites from widely separated terranes.

in Ti, Fe, and Mg in the biotite. The detailed color variations in the various specimens of biotite are listed in Tables 13 and 19, along with content of TiO_2, total Fe (as FeO), and MgO. Two very approximate correlations of color and composition of biotite are attempted. One is color versus TiO_2/FeO in the biotite, the other is color versus the ratio $(TiO_2:MgO)/(\text{total Fe(as FeO)})$ (Table 19). The correlation of color with variations of Ti, Mg, and Fe is essentially that suggested by Hall (1941).

Table 19 indicates the marked changes in the above-noted chemical ratios, and in color of biotites, as the paragneiss is traced from Group 1 at Emeryville, northeast to the massif near Colton (Group 5). In the least altered gneiss at Emeryville more than 80 per cent of the biotites are greenish brown to greenish black (Table 13). The ratio of TiO_2 to total Fe (as FeO) is on the order of 0.16. As least altered gneiss is traced northeast to Colton, this ratio changes to about 0.30. The successive changes in color are from greenish brown or greenish black through shades of reddish brown, very deep reddish brown, very deep red, to reddish black. The value of the relation $(TiO_2:MgO)/(\text{total Fe (as FeO)})$ over the simple ratio $TiO_2/(\text{total Fe (as FeO)})$ appears if comparisons are made of the differences in color between Bgn 21 and Bgn 19, or between Bgn 4 and Bgn 25, or between Qb 236 and Qb 228 (Table 13). Magnesia in relatively large amounts tends to mask or dilute the primary effects of Ti and Fe on color.

In most other metamorphic terranes where colors of biotites are recorded, the changes are

consistent with those we find—the biotite changes from greenish brown to red or reddish brown with increasing degree of metamorphism (Ambrose, 1936, p. 257; Harker, 1932, p. 214–217; Phillips, 1930; Sugi, 1935, p. 138–143; Tilley, 1925). Whether the ratio TiO_2:FeO in biotites increases and the total Ti content in the host rock decreases with increasing metamorphism in these terranes as in the paragneiss (Engel and Engel, 1958a, p. 1396) is unknown. Many examples of change in color of biotites from greenish brown to red or reddish brown with progressive metamorphism involve the appearance of almanditic garnet. In the partitioning of Fe and Mg between garnet and biotite, Fe almost invariably decreases in the biotite relative to Mg and probably Ti. The change to brown and reddish brown in many biotites may be influenced in part by this fractionation of elements between biotite and garnet.

TABLE 20.—POLYMORPHIC FORMS OF BIOTITE MICA IN THE PARAGNEISS

Powder-diffraction studies by H. S. Yoder, Jr.

Area	Sample	Polymorph
Emeryville (Group 1)	Qb 236	1 M or 3 T
	Qb 235 M	1 M or 3 T, and 2 M
	Qb 231	2 M
	Qb 216	1 M or 3 T, and 2 M
Russell (Group 3)	Bgn 6	1 M or 3 T
	Bgn 9	2 M
	Bgn 10	1 M or 3 T
Colton (Group 5)	Bgn 18	1 M or 3 T, and 2 M
	Bgn 21	1 M or 3 T
	Bgn 19	1 M or 3 T

Crystal Structure

Introduction.—Studies of crystal structure of the biotites indicate the presence of 1M, 2M, and possibly 3T polymorphs, but there is no clear correlation of polymorphic type and grade of metamorphism. The 1M polymorphs seem to predominate at both Emeryville and Colton. A small decrease probably occurs in the dimensions of the unit cell of biotites with increasing grade of metamorphism. There is no good correlation of Fe:Mg in the biotites with the intensity ratio of (004) to (005). This may be due principally to the large amounts of octahedrally co-ordinated Al in the biotites. Correlation of the composition of the biotites with crystal structure indicates a small deficiency of the large cations Ca, Na, and K and the possibility that oxygen substitutes for some OH or F (oxymicas?).

TABLE 21.—MEASUREMENTS OF THE REFLECTIONS (004) AND (005) AND CALCULATED d IN Å FOR BIOTITES FROM THE PARAGNEISS

Area	Sample	(004) reflection		(005) reflection	
		C sin B	d in Å	C sin B	d in Å
		$\bar{x}$*	$\bar{x}$*	$\bar{x}$*	$\bar{x}$*
Emeryville (Group 1)	Qb 227	10.050	2.5126	10.057	2.0116
	Qb 226	10.050	2.5127	10.058	2.0117
	Qb 230	10.047	2.5118	10.053	2.0107
	Qb 7	10.044	2.5114	10.052	2.0105
	$\bar{x}$†	10.047	2.5121	10.055	2.0111
Colton (Group 5)	Bgn 18	10.036	2.5090	10.041	2.0082
	Bgn 19	10.028	2.5072	10.033	2.0067
	Bgn 20	10.019	2.5047	10.026	2.0052
	Bgn 21	10.020	2.5052	10.028	2.0056
	$\bar{x}$†	10.026	2.5062	10.032	2.0064

* Mean of three separate runs in which the maximum mean deviation for C sin B is .0022 and for d in Å is .0005

† Mean of values for the four specimens in this group

Polymorphic types.—Powder-diffraction studies of 10 of the biotites from the paragneiss suggest that 1M polymorphs are widespread, and the 2M mica is not uncommon (Table 20). Although powder-diffraction studies do not differentiate 1M from the 3T micas, the 1M form appears to be at least 10 times more abundant in nature than the 3T (Hendricks and Jefferson, 1939; Levinson and Heinrich, 1954; Yoder and Eugster, 1954). We assume that this relation holds in the paragneiss. Yoder and Eugster (1954, p. 159) noted that the 1M mica appears in rocks formed throughout the entire thermal range of geological conditions, from sediments to lavas.

The 2M polymorphs definitely occur in specimens from Emeryville and Russell. Further X-ray studies probably will identify the 2M form at Edwards and Colton also. The X-ray data at hand from the paragneiss, and data reported by Hurst (1957), do not indicate any correlation of polymorphic type with grade or temperature of metamorphism.

A single flake of biotite from the gneiss (Qb

228, Emeryville) X-rayed by A. A. Levinson proved to be a 2M mica, but the powder-diffraction studies indicate that some homogeneous looking specimens of gneiss contain at least two polymorphic forms of biotite.

TABLE 22.—COMPARISONS OF THE RATIOS OF OBSERVED INTENSITIES OF (004) TO (005) REFLECTIONS IN BIOTITES IN THE PARAGNEISS WITH PREDICTED AND DETERMINED AMOUNTS OF Fe IN OCTAHEDRAL SITES

Area	Sample	I(004)/I(005)			Per cent of octahedral sites occupied by Fe	
		Max.	Min.	Mean*	Predicted (Gower, 1957)	Chem. anal.
Emery-	Qb 227	0.954	0.833	0.897	53.5	42.0
ville	Qb 226	0.850	0.820	0.836	49.0	42.0
(Group 1)	Qb 230	1.014	0.975	0.991	59.5	45.0
	Qb 7	1.162	1.032	1.075	64.0	45.0
	Mean†	0.995	0.915	0.950	56.5	44.0
Colton	Bgn 18	0.855	0.796	0.835	49.0	33.0
(Group 5)	Bgn 20	0.829	0.777	0.804	47.0	n.d.
	Bgn 21	0.992	0.845	0.927	55.0	36.0
	Bgn 19	0.823	0.755	0.796	46.5	32.0
	Mean†	0.877	0.793	0.841	49.4	33.7

* Mean of six runs

† Mean of values for the four specimens in this group

Unit cells and ratios of reflections (004) to (005).—Measurements of (004) and (005) reflections and dimensions of the unit cell are in Table 21. There appear to be real, but small differences in $C \sin \beta$ and d in Å between biotites at Emeryville and Colton. At least, these differences appear greater than the errors involved in the measurements.

If a decrease in size of unit cell of biotites occurs with increasing grade of metamorphism the change may be related to changes in composition of the biotites. It is clearly established that increases in F are accompanied by a marked decrease in dimensions of the unit cell (Gower, 1957; Yoder and Eugster, 1954), and that the F content of biotites in the paragneiss increases between Emeryville and Colton (Tables 13, 15).

The effects of pressure on cell dimensions are not positively resolved, but Crowley and Roy (1958) believe that the cell dimensions of synthetic micas are a reproducible function of the pressure of formation. The differences they record are 0.04 Å, between mica (muscovite) formed at 6000 psi and that formed at 60,000 psi.

The systematic change in the composition of the biotites with increasing grade of metamorphism may be reflected in various changes in crystal structure. The intensities of the reflections (004) and (005) have been studied by Gower (1957) who has attempted to correlate the ratio of the observed intensities of the (004) and (005) reflections with the ratio Fe:Mg and the per cent Fe in octahedral sites of the biotites. In an attempt to check Gower's method, the (004) and (005) reflections were measured on 8 biotites, 4 from Emeryville and 4 from Colton. The ratios of the (004)–(005) reflections are given in Table 22, along with the per cent of octahedral sites occupied by Fe and the actual amounts as determined by chemical analysis.

It is clear that the ratio of the observed intensities of (004) and (005) reflections decreases from 0.95 at Emeryville to 0.84 at Colton, as do both the predicted and chemically determined per cent of Fe in octahedral sites. (*See also* Tables 16 and 17.) Gower recognized that the method he proposed and his working curve were limited to biotites that contain principally magnesium and iron in the octahedral positions. This obviously is not applicable to the biotites in the paragneiss, which have from 15 to almost 30 per cent Al and Ti in octahedral positions (Table 16).

GARNET

Introduction

The various geological environments in the paragneiss complex are distinctly reflected in compositions of constituent garnets, but there is no correlation between composition and color. Index of refraction, crystal structure, and densities change slightly with the composition of the system and grade of metamorphism, but these changes are close to the errors of measurement. Garnets formed at contacts of paragneiss and pegmatite in the Emeryville area are almanditic (pyralspites) with 5 to 7 weight per cent MnO, 4 per cent MgO, and 2 per cent CaO. The first garnets to appear throughout the least altered gneiss (at Edwards) contain 1 to 1.5 per cent MnO, 5.5 to 6.5 per cent MgO, and about 1.5 per cent CaO. As the grade of metamorphism increases from Edwards to Colton, garnets in least altered gneiss are enriched in MgO

TABLE 23.—REPRODUCIBILITY OF CHEMICAL ANALYSES OF MAJOR CONSTITUENTS IN GARNETS

Oxide	Qb 226						Bgn 18			
	1	2	3	Mean	Max. deviation	Mean deviation	1	2	Mean	Mean deviation
SiO_2	37.72	37.88	37.92	37.84	.12	.08	38.82	38.54	38.68	.14
TiO_2	n.d	.02	.02				.03	.02	.03	.01
Al_2O_3	20.87	21.00	21.12	21.00	.13	.09	21.94	21.87	21.91	.04
Fe_2O_3	2.32	1.94	2.13	2.13	.19	.13	.67	.94	.81	.14
FeO	25.45	25.52	25.31	25.43	.12	.08	29.36	28.99	29.18	.18
MnO	6.84	6.88	6.75	6.82	.07	.05	.70	.72	.71	.01
MgO	4.64	4.71	4.87	4.74	.13	.09	7.68	7.64	7.66	.02
CaO	2.36	2.52	2.46	2.45	.09	.06	1.20	1.22	1.21	.01

Qb 226, number 1 and Bgn 18, number 1; analyst, L. M. Tarrant
Qb 226, numbers 2 and 3, and Bgn 18, number 2; analyst, C. G. Engel

TABLE 24.—COMPARISON OF MnO DETERMINATIONS IN GARNETS, USING THE SPECTROPHOTOMETER AND EMISSION SPECTROGRAPH

Sample	Spectrophotometer	Emission spectrograph
Bgn 19	0.55	0.52
Bgn 21	0.58	0.57
Bgn 18	0.72	0.67
Bgn 23	1.32	1.29
Bgn 22	1.13	0.64
Bgn 16	0.92	0.64
Bgn 6	1.40	1.35
Bgn 10	1.55	1.42
Bgn 9	0.96	1.10
Bgn 12	1.18	0.98
Bgn 13	0.83	0.64
Qb 100	0.97	1.03
Qb 233	1.21	1.29
Qb 226	6.82	6.45

(up to 7.7 per cent) and impoverished in MnO (0.64 per cent). These changes in composition of the garnets may be accompanied by a decrease in density of about 0.05 and by slight changes in the unit cell. They are not reflected in indices of refraction or color.

Ten garnets have been analyzed *in toto*, 2 each from Groups 1, 2, 3, and 5, 1 from Group 4, and 1 from the paragneiss between Groups 3 and 4 (Table 25). Partial analyses have been made on an additional 10 garnets (Table 26). These analyses are supplemented by determinations of density, index of refraction, and X-ray measurements of the unit cell (Table 28).

The dominant changes in composition of the garnets are plotted against grade of metamorphism in Figures 8 and 9. The relationships of composition of the garnets in the paragneiss to compositions of garnets reported from other biotite-bearing gneisses and schists are shown in Figure 10 (after Wright, 1938). A photomicrograph of typical garnet in the least altered gneiss at Colton appears in Plate 3.

Analytical Methods

Analyses of major elements.—The 10 analyses of major elements in garnet are reported (Table 25). Specimen Qb 226 was analyzed 3 times, once by Lee C. Peck and twice by C. G. Engel; and Bgn 18 was analyzed by both C. G. Engel and L. M. Tarrant (Table 23). All other analyses of major and minor constituents in Tables 25 and 26 are by C. G. Engel. Differences in composition of either Qb 226 or Bgn 18, resulting from work of different analysts, are small compared to differences in the composition of garnets of different groups. The MnO content of most garnets was determined both by spectrophotometer and independently by emission spectrograph. The results of these analyses (Table 24) indicate the extent of agreement between these two analytical techniques.

Errors introduced by impurities in reported compositions of garnets are small because of the lack of alteration of the garnets and the relative ease with which they separate from most asso-

ciated minerals. All garnet samples are 98 per cent or more pure. Quartz is the only impurity in excess of 0.5 per cent. In Group 5, at Colton, quartz composes as much as 0.2 to 1 per cent of the analyzed samples. The difficulty of separating out all quartz is due to the extremely intricate way it sieves the garnet grains. Analyses of garnet by Peck, Tarrant, and Engel employ classical methods.

Index of refraction.—Indices of refraction of the garnets are given in Table 25, below the chemical compositions of each specimen. Four indices of refraction (N) are recorded for each specimen. These four values of N were obtained (1) by conventional oil-immersion techniques with sodium light, (2) from the indices of refraction of end members as determined on naturally occurring garnets by Ford (1915), (3) from the indices of refraction of synthetic end members as determined by Skinner (1956, p. 429), and (4) from the relation $(dK + 1)$ where $(n-1)/d = K$[11]. In most specimens the first three methods of determination of N agree within .002 which is within the error of measurement, and the maximum difference in N by these three methods is only .004 (Qb 228, Table 25). In contrast, the calculations of N from the rule of Gladstone and Dale vary from the determined N by more than .004 in all but 2 specimens. Five of the determinations by the rule of Gladstone and Dale deviate from the determined N by .01. The deviations are in the same direction in all instances, and 7 out of 10 differences range between .007 and .01. This wide, fairly systematic deviation in the value of N reflects either (1) inaccuracies in optical, chemical, or density data, or (2) the inadequacy of the rule of Gladstone and Dale. Recently Jaffe (1956) reinvestigated the applicability of the relation $(n-1)/d = K$, using the data for 121 minerals selected from modern literature. Jaffe (1956, p. 757) concludes:

> "... where the optical, chemical and density data are accurate, the Gladstone and Dale relations hold very well for most minerals, with only slight modifications due to the manner of combination of the constituents. Mineralogists would do well to reexamine their data before deciding that a mineral is an exception to the rule $(n - 1)/d = K$".

If we must mistrust our data[12], it would seem that the least accurate measurements are of density, as noted in the following section.

The determination of N in immersion oils, or from either naturally occurring or synthetic end members (methods 1, 2, and 3, Table 25) point up one important fact. No large, systematic variations in N are induced by the appreciable differences in chemical composition of garnets such as Qb 226 and Bgn 25 or Bgn 15 (Fig. 8; Table 25). Stated conversely, determination of N alone is of little use in determining the chemical composition of many garnets, or of many micas or amphiboles for that matter.

Certainly Barth (1936, p. 786) does not appear justified in stating:

> "For all [almanditic] garnets [in Dutchess County, New York] a value of 1.810 ± 0.005 seems to be correct. . . .This indicates that all garnets of the present area have essentially the same composition, regardless of metamorphic grade".

Density.—Three values for density of most garnets are listed in Table 25. These values were obtained by (1) direct determinations using a pycnometer, (2) calculations from Fleischer's data on naturally occurring garnets (Fleischer, 1937), and (3) calculations from Skinner's measurements of the density of synthetic end members (Skinner, 1956, p. 430).

All three of these methods agree within the limits of reproducibility of our measurements (±.04) and those of Skinner (±.03), although it obviously would be desirable to refine the measurements.

Occurrence

Garnetiferous migmatite, Emeryville area.—Garnet appears in the Emeryville area where pegmatites and granitic veins cut the gneiss, but to the northeast it occurs throughout least altered gneiss as a product of progressive regional metamorphism. The close association in the Emeryville area of garnets with injected, pegmatitic gneiss has been discussed in Part I of these studies (1958a, p. 1395) and earlier (Engel and Engel, 1953, p. 1072–1076). Typically the garnets are in or scattered along a cluster of quartz which is in turn enveloped in a schliere of gneiss in pegmatite or bordered on one side by gneiss and on the other by quartz and K feldspar veinlets or pegmatite. Associated biotite is commonly reddish rather than greenish brown, and the K feldspar content of the gneiss commonly exceeds 5 per cent (Engel and Engel, 1953, Fig. 10, p. 1073). Chlorite, replacing biotite, is more abundant in the garnetiferous mig-

[11] The rule of Gladstone and Dale (*see* Larsen and Berman, 1934; Jaffe, 1956, p. 757–761), in which d = density, K = the specific refractive energy of a chemical compound, generally an oxide.

[12] Calculations of N (by the rule of Gladstone and Dale) for most almandites cited in the literature result in large discrepancies.

TABLE 25.—CHEMICAL ANALYSES, MOLECULAR COMPOSITION, INDEX OF REFRACTION, AND DENSITY OF GARNETS FROM THE PARAGNEISS IN THE EMERYVILLE-COLTON REGION, NEW YORK

	Emeryville Group 1			Edwards Group 2			Russell Group 3			South Pierrepont Group 4			Colton Group 5		
	Qb 226	Qb 228	$\bar{x}$	Qb 233	Qb 100	$\bar{x}$	Bgn 9	Bgn 14	$\bar{x}$	Bgn 15	Bgn 25	$\bar{x}$	Bgn 18	Bgn 21	$\bar{x}$
SiO_2	37.72	37.50	37.61	37.96	38.03	37.99	37.64	38.30	37.97	38.06	38.10	38.08	38.82	38.44	38.63
Al_2O_3	20.87	21.35	21.11	22.13	22.05	22.09	22.62	22.45	22.52	22.03	23.49	22.76	21.94	22.61	22.28
Fe_2O_3	2.32	1.57	1.95	1.05	.88	.97	1.66	1.04	1.35	.99	1.29	1.14	.67	.85	.76
FeO	25.45	29.60	27.53	31.14	29.17	30.16	30.64	29.39	30.02	31.30	29.18	30.24	29.36	28.62	28.98
MnO	6.84	5.12	5.98	1.21	1.57	1.39	.96	1.36	1.16	.70	.46	.58	.70	.58	.64
MgO	4.64	3.73	4.19	5.59	6.49	6.04	5.18	5.15	5.17	6.07	6.28	6.11	7.68	7.66	7.67
CaO	2.36	1.24	1.80	1.32	1.80	1.56	1.38	2.60	1.99	.72	1.60	1.16	1.20	1.62	1.41
Total	100.20	100.11		100.40	99.99		100.08	100.29		99.87	100.40		100.37	100.38	
Trace Elements (In parts per million)															
Ba	10	7	8	7	3	5	10	15	5*	10	..	..*	2	2	2*
Co	9	6	8	30	25	28	20	18	13	15	22	22	30	32	22
Cr	18	12	15	30	45	38	40	55	37	35	44	41	44	68	52
Cu	2	2	2	34	25	29	6	2	..	3	6	..	2	10	..
Ni	tr	23	..	27	35	31	tr	..	..	..	10	..	..	tr	..
Sc	550	800	675	450	550	500	500	450	442	350	250	283	250	300	214
Sn	16	tr	..	..	..	..	..	..	..	..	..	..	..	..	..
Ti	135	125	130	210	120	165	150	280	179	300	200	233	150	290	241
V	25	18	21	29	30	30	31	30	29	40	33	38	35	40	39
Y	2000	1500	1750	1500	1100	1300	1200	1300	1300	1000	900	900	550	500	679
Yb	520	460	490	500	300	400	340	300	265	80	100	87	36	35	75
Zn	75	140	107	130	200	165	200	100	162	210	225	198	275	280	235
Zr	160	130	145	250	160	205	150	135	113	200	200	160	150	180	114
Molecular Composition†															
Almandite	60.2	69.9	65.0	71.5	66.0	68.8	72.8	68.6	70.7	71.1	69.0	69.0	65.0	64.2	64.6
Spessartite	15.8	12.2	14.0	2.8	3.6	3.2	2.2	3.2	2.7	1.6	1.1	1.3	1.6	1.3	1.5
Pyrope	19.1	15.7	17.4	22.9	26.2	24.6	21.0	21.2	21.1	24.8	25.9	25.2	31.4	31.5	31.4
Grossularite	4.9	2.2	3.6	2.8	4.2	3.5	4.0	7.0	6.1	2.5	4.0	4.5	2.0	3.0	2.5
Index of Refraction															
n_D	1.815	1.830		1.820	1.820		1.820	1.820		1.820	1.820		1.815	1.815	
Density															
d_{25}^{25}	4.19	4.23		4.19	4.19		4.19	4.19		4.19	4.19		4.15	4.15	

*Not determined.
†Molecular proportions recalculated to 100 percent.

Zn	75	140	107	130	200	165	200	100	162	210	225	198	275	280	235
Zr	160	130	145	250	160	205	150	135	113	200	200	160	150	180	114
Molecular Composition†															
Almandite	60.2	69.9	65.0	71.5	66.0	68.8	72.8	68.6	70.7	71.1	69.0	69.0	65.0	64.2	64.6
Pyrope	18.1	15.0	16.5	22.2	25.5	23.8	20.9	20.8	20.8	24.5	25.5	25.5	30.1	29.9	30.0
Grossularite	6.6	3.4	5.0	3.8	5.0	4.4	4.0	7.5	5.8	2.1	4.7	4.7	3.3	4.6	4.0
Spessartite	15.1	11.7	13.5	2.5	3.5	3.0	2.2	3.1	2.7	2.3	.8	.8	1.6	1.3	1.4
Index of Refraction (N)															
Measured	1.797	1.802	1.800	1.801	1.793	1.797	1.797	1.797	1.797	1.796	1.798	1.797	1.793	1.791	1.792
Using Ford's data	1.797	1.804	1.801	1.798	1.792	1.795	1.797	1.797	1.797	1.797	1.797	1.797	1.793	1.789	1.791
Using Skinner's data	1.789	1.806	1.802	1.800	1.794	1.797	1.799	1.798	1.799	1.799	1.799	1.799	1.796	1.791	1.794
Rule of Gladstone and Dale	1.810	1.820	1.815	1.812	1.795	1.804	1.813	1.806	1.810	1.810	1.805	1.808	1.802	1.801	1.802
Density															
Determined	4.10	4.12	4.11	4.12	4.08	4.10	4.12	4.08	4.10	4.11	4.08	4.10	4.06	4.05	4.06
Using Fleischer's data	4.10	4.16	4.13	4.11	4.07	4.09	4.12	4.09	4.11	4.11	4.08	4.10	4.05	4.04	4.05
Using Skinner's data	4.12	4.18	4.15	4.12	4.09	4.11	4.13	4.12	4.13	4.12	4.10	4.11	4.07	4.06	4.07

* Columns include spectrographic analyses of garnets in Table 26.

† All Fe calculated as Fe^{++}

Analysts (major elements): Bgn 18, Bgn 15, L. M. Tarrant; Qb 226, Lee C. Peck; Bgn 9, Bgn 14, Bgn 21, Bgn 25, Qb 100, Qb 228, Qb 233, Celeste G. Enge

TABLE 26.—TRACE-ELEMENT COMPOSITION OF GARNETS NOT ANALYZED *in toto*

Element	Russell (Group 3)				South Pierrepont (Group 4)	Colton (Group 5)				
	Bgn 13	Bgn 12	Bgn 6	Bgn 10	Bgn 16	Bgn 23	Bgn 22	Bgn 24	Bgn 20	Bgn 19
Ba	..	2	2	2	..	2	2	2	2	2
Co	12	12	8	15	30	14	14	20	20	25
Cr	30	35	34	25	45	55	40	40	55	6
Mn	4000	6000	7000	8400	3000	8000	4400	5000	3200	3300
Ni	tr	23	tr	..	5?	tr	..	tr	10	..
Sc	350	450	450	450	250	250	150	200	150	200
Ti	190	190	110	155	200	300	300	250	200	200
V	20	30	30	30	40	35	40	35	45	45
Y	600	1200	1500	2000	800	1000	800	800	500	600
Yb	200	250	300	200	80	190	80	110	36	40
Zn	200	120	175	175	160	140	250	200	250	250
Zr	75	120	130	70	80	135	75	110	50	100

matite than in most of the less injected paragneiss.

The characteristic occurrence of garnet at or near contacts of pegmatite and gneiss suggests that garnet is a product of reciprocal reaction between constituents of these very different rocks. As such, this garnet may not have formed in equilibrium with the biotite of the gneiss, and the Emeryville garnets must be distinguished from garnets northeast of Edwards that occur within the body of little injected gneiss[13].

The development of the garnet at and near pegmatite contacts and its absence in the least altered gneiss near Emeryville is perhaps best explained by the Mn content of the garnets (Tables 25, 26). The Emeryville garnets contain four times as much MnO as those in the least altered gneiss at Edwards, and Mn-rich almandite garnets are stable at much lower grade of metamorphism than Mn-free garnets. This characteristic of Mn greatly to expand the stability field of garnets was indicated by the early work of Goldschmidt (1921) at Stavanger, and expanded by Tilley (1925). It is now well documented by extensive evidence (Ramberg, 1949; Miyashiro, 1953).

Characteristics of the pegmatite–gneiss contact in the Emeryville area other than Mn content may have enhanced the development of garnet. The water content of the system may have been especially influential. The role of water, relative to T, in metamorphic systems has been discussed by Yoder (1955, p. 521), extrapolating from experimental studies. Yoder notes that garnet may form at pegmatite–gneiss contacts if:

"(1) . . . the contact is dry, and therefore garnet would be preserved at any temperature below its breakdown curve. (2) The water content of the rocks is so low that their bulk composition lies in the water deficient region, and hence garnet is stable with hydrous minerals in the absence of free water at temperatures below the breakdown temperature of the hydrous minerals".

Garnet in least altered gneiss.—The appearance of the garnet in least altered gneiss at Edwards coincides with the disappearance of muscovite and the decrease in coexisting biotite. These reactions, with increasing degree of metamorphism, are commonly illustrated by use of the equation: biotite + muscovite + quartz + magnetite → almandite + water + orthoclase.

Actually, appearance of garnet in this rock at Edwards is accompanied by loss of H_2O, SiO_2, and K_2O (Part I, Engel and Engel, 1958a, p. 1396–1399), and the amount of orthoclase formed in the paragneiss as a product of this reaction at Edwards is negligible. Because of the expulsion of K and H_2O, the formation of garnet in the paragneiss at Edwards is perhaps more realistically expressed as: biotite + muscovite + quartz + magnetite→ almandite and loss of K_2O, SiO_2 and H_2O from the rock[14].

[13] For this reason in Figure 8, which compares physical and chemical properties of garnets in the Emeryville area with properties of garnets to the northeast, the compositional curves are broken between Emeryville and Edwards by queries.

[14] Sometimes the use of equations to define metamorphic reactions is more misleading than helpful,

TABLE 27.—COMPOSITIONS OF ALMANDITIC GARNETS RELATED TO GRADE OF METAMORPHISM

Oxide	Locality	Amphibolite facies*		Granulite Facies
		Upper middle	Upper	Lower
SiO_2	Abukuma Plateau, Japan†		37.66	
	Adirondack paragneiss**	37.61	37.99	38.63
	Dutchess County, New York††	37.94	37.58	
	Aberdeenshire***		36.59	
Al_2O_3	Abukuma Plateau		20.30	
	Adirondack paragneiss	21.11	22.09	22.28
	Dutchess County	22.87	21.58	
	Aberdeenshire		22.42	
Fe_2O_3	Abukuma Plateau		1.75	
	Adirondack paragneiss	1.95	.97	.76
	Dutchess County		n.d	
	Aberdeenshire		.04	
FeO	Abukuma Plateau	26.35	29.62	
	Adirondack paragneiss	27.53	30.16	28.98
	Dutchess County	32.45	36.69	
	Aberdeenshire		32.11	
MnO	Abukuma Plateau	5.75	2.87	
	Adirondack paragneiss	5.98	1.39	.64
	Dutchess County	1.77	n.d	
	Aberdeenshire		1.42	
MgO	Abukuma Plateau		3.88	
	Adirondack paragneiss	4.19	6.04	7.67
	Dutchess County	1.83	2.31	
	Aberdeenshire		5.41	
CaO	Abukuma Plateau		2.50	
	Adirondack paragneiss	1.80	1.56	1.41
	Dutchess County	2.04	1.89	
	Aberdeenshire		.54	

* Correlations based upon mineralogy of host rocks
† Miyashiro (1953)
** This study
†† Barth (1369)
*** Stewart (1950)

In the Russell-Colton area, increasing amounts of garnet are formed as biotite is resorbed, and there is continued loss of Si and K from the rock (Fig. 1). The sum of modal garnet plus biotite at Colton is roughly that of the modal biotite in the least altered gneiss at Emeryville.

especially if detailed chemical compositions of the minerals are unknown. Equations involving only some of the major constituent minerals in the rock are rarely even approximately correct. In the instance of the paragneiss, this is indicated from inspection of the compositions of the constituent minerals, from the fact that the bulk composition of the system is changed, and from field relations which indicate a decrease in volume of the total system. Metamorphic reactions must be considered in the context of the actual composition and amounts of all the minerals and the behavior of the gross system rather than in terms of an isolated group of minerals of fictitious (idealized) compositions.

Composition Related to Metamorphic Grade

The garnets in the paragneiss became richer in MgO, Cr, Ti, and Zn and depleted in MnO, Fe_2O_3, Yb, and Y with increasing grade of metamorphism. The content of Sc probably decreases, and FeO may decrease. Amounts of V may increase with increasing metamorphic grade. These conclusions are based upon the analyses listed in Tables 25 and 26 and illustrated in graphical form in Figure 8.

TABLE 28.—DIMENSIONS OF UNIT CELLS OF GARNETS AS DETERMINED FROM (420) AND (642) REFLECTIONS, AND AS CALCULATED FROM SKINNER'S DATA ON SYNTHETIC END MEMBERS

Area	Specimen	Unit cell (a_o)		a_o after Skinner*
		(420) 1	(642) 2	3
Emeryville (Group 1)	Qb 226	11.528	11.530	11.549
Edwards (Group 2)	Qb 233	11.530	11.533	11.525
	Qb 100	11.529	11.528	11.529
Colton (Group 5)	Bgn 18	11.522	11.521	11.518
	Bgn 21	11.524	11.525	11.522
	Bgn 20	11.517	11.517	....
	Bgn 19	11.519	11.522	....

* a_o calculated from chemical analyses and a_o values of synthetic garnets by Skinner (1956)

Comparisons have been attempted with the few seemingly analogous data in the literature, but the results are far from satisfactory (Table 27). In the recorded examples where garnets coexisting with biotites in pelitic schists have been analyzed the grade of metamorphism does not seem to exceed that found at Edwards. This conclusion is necessarily tinged with considerable doubt because of the fragmentary reports of the composition of parent rocks and their geological setting. The work of Miyashiro (1953; 1958) on the garnet-bearing metasedimentary schists of Abukuma Plateau in Japan is an exception. Comparisons also are attempted between our data and those published by Barth (1936) on the garnets in the metamorphosed Paleozoic schists of Dutchess County, New York. The assumption is made in the groupings of garnet analyses in Table 27 that the highest-grade assemblages reported in pelitic schists by Miyashiro and Barth are crudely analogous in metamorphic facies of the Adirondack paragneiss at Edwards. The column headed upper-middle amphibolite facies (Table 27) includes garnets from (1) Miyashiro's Metamorphic Zones III and IV, (2) Bald Mountain (Clove quadrangle) as reported by Barth, and (3) the pegmatitic gneiss at Emeryville. In the middle column (upper amphibolite facies) are grouped garnets from (1) upper part of Zone IV Abukuma Plateau, the highest-grade assemblage reported there, (2) Redding, Connecticut, reported by Barth as from the same parent rock as the garnet of Bald Mountain, but representing a higher grade of metamorphism, and (3) the mean of the two garnets from the paragneiss at Edwards that have been analyzed in detail. The column labeled "lower granulite facies" contains only the mean of the two analyses of garnets from the paragneiss at Colton.

Probably even the comparisons in Table 27 are highly questionable, especially comparison between garnets in pegmatitic gneiss at Emeryville and those of the Japanese and Dutchess County specimens, which appear to be products of less additive regional metamorphism. But the crude trends reflected by the analyses appear to be fairly consistent, and the groupings appear to substantiate the compositional trends tentatively suggested in Figure 9.

In the following paragraphs the changes in composition of the garnets in the paragneiss are discussed briefly and related to Figure 9 and to the work of Myashiro, Barth, and others.

Silica.—The apparent increase in SiO_2 content of garnets at Colton (Group 5, Tables 25, 27) is probably due to quartz impurities. Several of the garnets in the gneiss, but especially Bgn 18 at Colton, are intimately sieved with quartz, which constitutes about 1 volume per cent of the analyzed sample. Probably most of the garnets between Edwards and Colton have essentially the same SiO_2 content as those at Emeryville. Most of the variations in SiO_2 and Al_2O_3 in analyses of garnet (*see especially* Wright, 1938) are probably due to impurities and to inadequate analytical techniques of the type discussed in Fairbairn *et al.* (1951).

Aluminum.—No real differences appear in the Al_2O_3 content of the garnets found within the body of the paragneiss, and there is no reason to expect differences. Probably the Al_2O_3 contents reported in the analyses by C. G. Engel are slightly high because of difficulties encountered in getting gas-burner temperatures high enough to satisfactorily "burn off" the large R_2O_3 group.

Iron.—Total iron content and the amounts of ferrous and ferric iron in the garnets show in-

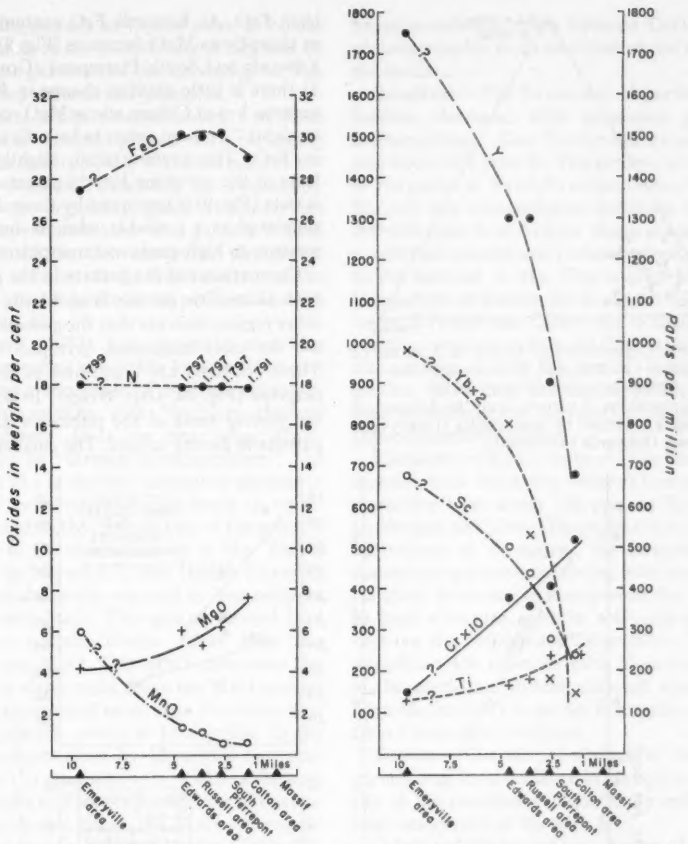

FIGURE 8.—VARIATIONS IN CHEMICAL COMPOSITION AND INDEX OF REFRACTION OF GARNETS PLOTTED AS A FUNCTION OF INCREASING GRADE OF METAMORPHISM BETWEEN EMERYVILLE AND COLTON, NEW YORK

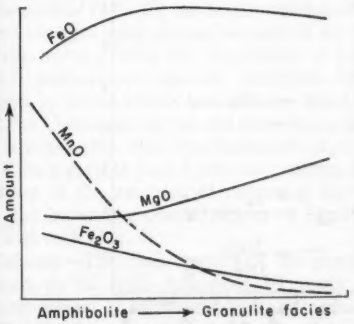

FIGURE 9.—TENTATIVE INTERPRETATION OF VARIATIONS IN COMPOSITION OF ALMANDITIC GARNETS IN BIOTITE-BEARING SCHISTS AND GNEISSES METAMORPHOSED TO UPPER AMPHIBOLITE —LOWER GRANULITE FACIES

teresting variations related to the composition of the rock and its grade of metamorphism (Figs. 8, 9, 11). Small amounts of ferric iron seemingly exist in all the garnets.[15] The amount of ferric iron appears to be highest in the more spessartitic almandites in the pegmatitic gneiss at Emeryville (Table 25). Presumably the Fe_2O_3 content of garnets found within the body of the gneiss, between Edwards and Colton, decreases with increasing rank of metamorphism. This decrease is sympathetic with the decrease in MnO content (Fig. 9). The variations in FeO appear to be real and more complicated. The spessartitic garnets at Emeryville contain the

[15] Because of the very small amounts of Fe_2O_3, especially in the garnets in Groups 2 to 5, the Fe^{+++} was converted to Fe^{++} in all calculations of garnet "end members".

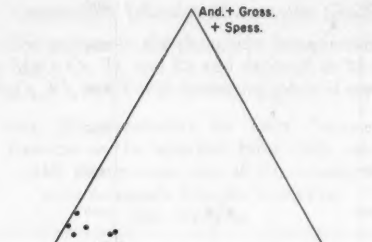

FIGURE 10—COMPOSITIONAL RANGE OF ALMANDITIC GARNETS IN BIOTITIC PARASCHISTS AND GNEISSES (after Wright, 1938)
The composition of garnets from the Adirondack paragneiss is indicated by open circles (Emeryville) and crosses (Edwards to Colton).

least FeO. At Edwards FeO content increases as abruptly as MnO decreases (Fig. 8). Between Edwards and South Pierrepont (Groups 2 and 4) there is little obvious change in FeO in the garnets, but at Colton, where MgO content rises to about 7.7, there seems to be a slight decrease in FeO. The asymmetrical, slightly domical form of the curve for FeO in garnets of biotite schists (Fig. 9) is prompted by these data and is suggested as a probable relation during intermediate to high-grade metamorphism.

Comparisons of the garnets in the paragneiss with almandinitic garnets from biotite schists of other regions indicate that the garnets at Colton are the most magnesian (pyropic) on record. These relations are illustrated in the ternary diagram (Fig. 10, after Wright, 1938) in which are plotted most of the published analyses of garnets in biotite schists. The smaller amounts

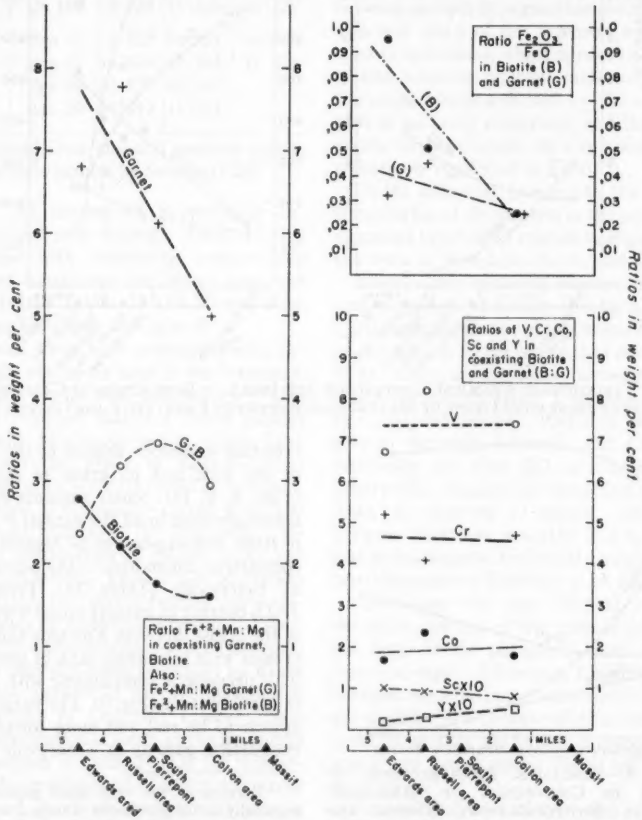

FIGURE 11.—FRACTIONATION OF ELEMENTS IN COEXISTING BIOTITE AND GARNET IN THE EMERYVILLE-COLTON REGION

of Fe in the garnets at Colton seem to be the natural result of the partitioning of Mg and Fe between the garnets and the coexisting biotites at high temperatures and pressures.

Manganese.—The marked decrease in MnO content in almanditic garnets with increasing grade of metamorphism is well illustrated in the specimens studied. Presumably the availability of Mn in the pegmatitic gneiss at Emeryville was an important reason why garnets formed there and not in the uninjected paragneiss. Northeast of Edwards the MnO content of the garnets decreases to about 0.6 per cent MnO at Colton (Fig. 8).

Miyashiro (1953) has suggested that the highest concentrations of Mn in almanditic garnets of any given grade of metamorphism appear in conjunction with "ordinary thermal metamorphism." It is not clear what Miyashiro means by "thermal metamorphism." The amount of Mn in the first almanditic garnets to appear in metamorphic rocks seems to be directly related to the composition of the system, especially to the concentrations of Mn. This is borne out by Miyashiro's data (1953). The most spessartitic almandite reported by him contains 19.7 per cent MnO. This garnet formed in a quartz-biotite-albite-chlorite schist that has 0.41 per cent MnO. This MnO content in the host rock is about eight times the MnO content of the paragneiss and more than five times that of the pegmatitic gneiss at Emeryville. In the paragneisses described by Miyashiro that contain about the same amounts of Mn as the pegmatitic gneiss at Emeryville, and are of equivalent metamorphic grade, the MnO contents of constituent garnets are very similar (Table 27).

Magnesium.—The increase in MgO with increasing grade of metamorphism indicated by the analyses (Table 25) has been verified by different analysts using masked samples of identical specimens (Table 23). Inspection of published analyses of garnets coexisting with biotites in pelitic schists and gneisses indicates that the Colton garnets are the most magnesian yet reported (Fig. 10). The increase in MgO in the Colton garnets (to 7.7 per cent) reflects the decrease in the amounts of coexisting biotite and the increasing concentrations of MgO in the host rock (Fig. 1).

Calcium.—The CaO content of the garnets appears to be little influenced by grade of metamorphism or by the moderate increments of CaO in the rock at higher grade of metamorphism (Table 25; Fig. 1). It is not clear from the published data whether this conclusion is generally true in pelitic schists. Certainly a broad correlation exists between CaO content of metamorphic rocks and that of the constituent garnet.

Scandium.—The Sc content of garnets, as in biotites, decreases with increasing grade of metamorphism. The Emeryville garnets contain about 675 ppm Sc. The garnets in the body of the gneiss at Edwards contain some 500 ppm Sc, and this concentration decreases to about 200–225 ppm Sc at Colton. These changes in Sc content are parallel and probably closely related to the decrease in Mn. The ratio of Mn:Sc in the garnets at Emeryville is almost 80, but between Edwards and Colton the ratio Mn:Sc is roughly constant at 20 to 23. The ratio of Y:Sc also remains crudely the same (3.0–4.0) in the garnets throughout the region, as might have been predicted from the marked coherence of these elements in nature.

Titanium.—The amounts of Ti in the garnets increase with increasing temperature of metamorphism from about 130 ppm at Emeryville to 240 ppm at Colton. These relatively low concentrations of Ti suggest the incapability of almanditic garnets—coexisting with biotite and traces of ilmenite—to incorporate this element in their structure, even in a Ti-rich province such as the Adirondack Mountains. There is more than 100 times as much Ti in biotites as in the coexisting garnets, although the ratio of Ti in biotite to Ti in garnet is roughly constant from Emeryville to Colton.

Barium.—The virtual absence of Ba in the garnets and its abundance in feldspars and biotite of the paragneiss undoubtedly reflects the close association of Ba with K.

Cobalt and Nickel.—About 8 ppm of Co is in the garnets at Emeryville. The Co content doubles or triples in the garnets between Emeryville and Edwards, but there is no systematic change in the content of Co with increasing grade of metamorphism between Edwards and Colton. The Ni content of the garnets is not clearly established, but probably there is about 10–20 ppm at Emeryville and between 5 and 35 ppm in the garnets to the northeast.

Chromium.—The garnets with relatively low Fe and Mg at Emeryville contain the smallest amounts of Cr (15 ppm). To the northeast between Edwards and Colton, Cr appears to increase slightly with increasing grade of metamorphism. The ratios of Cr:Mg and Cr:Ti in the garnets remain essentially constant, but there is no correlation between the distribution of Cr and either Al^{+++} or Fe^{+++} such as is thought to exist in some garnets.

Vanadium.—Amounts of V in the garnets

may increase slightly from 20 ppm at Emeryville to 40 ppm at Colton. If this increase is real it is opposite the trends of Fe^{++} and Fe^{+++} and parallel to that in Ti, with which V may be associated in some minerals. The ratio of V:Ti in the garnets remains approximately constant at .16 to .18 throughout the Edwards-Colton region.

Yttrium.—The Y in the garnets decreases with increasing grade of metamorphism from 1750 ppm at Emeryville to 680 ppm at Colton. The parallel trends of Y with Mn and Sc are consistent with observations on the chemical coherence of these elements.

Ytterbium.—The Yb content of the garnets is only slightly below the amounts of Sc and roughly one-fifth that of Y. Yb, like Sc, Mn, and Y, decreases in abundance with increasing grade of metamorphism of the garnets, but the decrease appears to be more extreme, from ∽490 ppm Yb at Emeryville to 75 ppm at Colton. The data suggest that the ratios Yb/Y and Yb/Sc in the garnets decrease with increasing rank of metamorphism.

Zinc.—The concentration of Zn in garnets increases between Emeryville and Colton from about 100 ppm to 235 ppm. This is parallel to the increase in Mg although by slightly larger increments, and partly opposite the trend in ferrous iron. The association of elements, if one exists, is probably between Zn and Mg. The Zn values in the garnets substantially bracket the concentrations of Zn reported for igneous rocks (Lundegårdh, 1947).

Zirconium.—The highest Zr values (∽200 ppm) in the garnets could reflect a trace of zircon, although no zircon is visible in the garnet concentrates we studied. No trend in amounts of Zr in the garnets can be related to geologic occurrence or degree of metamorphism.

X-Ray Studies

Dimensions of the unit cell (a_o) of seven of the garnets from the paragneiss have been determined (Table 28). One of the garnets is from the Emeryville area (Qb 226), two are from Edwards (Qb 233, Qb 100), and four are from Colton (Bgn 18, Bgn 19, Bgn 20, Bgn 21). The calculations of a_o are based on measurements of the reflections (420) and (642), with a maximum error of ±.0025 as noted in the section on analytical methods. Additional data for comparative purposes are: (1) a_o values for five of the garnets as calculated from the chemical analyses using Skinner's data (1956) for synthetic end members of the garnet group. Agreement between the directly determined values of a_o and those calculated from the chemical analyses and Skinner's X-ray data is good for three of the specimens but is poor for the most spessartitic garnet (Qb 226) from the Emeryville area.

Color

The slight differences in color of the garnets are not correlative with differences in composition, geologic occurrence, or grade of metamorphism. Most of the garnets are wine red, but the intensity of red varies from sample to sample, and a few garnets have a tinge of violet. In general the violet-tinged garnets are more translucent than the red garnets.

Partitioning of Elements Among Minerals

Introduction

General discussion.—Measurements of the selective distribution (fractionation) of elements among minerals of metamorphic rocks will contribute enormously to an understanding of their origin. These measurements, coupled with experimental studies, can indicate whether equilibrium is approached or achieved during crystal growth. Many distribution factors of elements are strongly temperature dependent, and the potential value of these factors in geothermometry is already apparent. Thoughtful and accurate studies of the selective distribution of elements among coexisting minerals will also permit a re-evaluation of Goldschmidt's classical concepts of capture, camouflage, and admittance—concepts based upon ionic characteristics, especially size and charge (Goldschmidt, 1954, p. 40–125). Presumably departures from ionic bonding toward more covalent types—as indicated by properties such as electronegativity—influence the distribution of elements, as may adsorption and other phenomena. These and a wide variety of processes and conditions related to the evolution and differentiation of the earth can be better understood when we know more about how the elements are distributed among metamorphic minerals.

Most metamorphic systems are so complicated, and our analyses of them sufficiently incomplete and inaccurate, that understanding comes slowly. It would be misleading to argue that our studies of the gneiss are a conspicuous exception. The purity of our samples and the accuracy and precision of our analyses fall short

Table 29.—Kinds of Data Available on Composition of Ten Specimens of Paragneiss and Their Constituent Minerals for which Fractionation Ratios are Calculated in Tables 31 and 32 and Figures 11 and 12

TABLE 29.—KINDS OF DATA AVAILABLE ON COMPOSITION OF TEN SPECIMENS OF PARAGNEISS AND THEIR CONSTITUENT MINERALS FOR WHICH FRACTIONATION RATIOS ARE CALCULATED IN TABLES 31 AND 32 AND FIGURES 11 AND 12

Area	Specimen number	Type paragneiss	Determinations, composition total rock			Composition constituent minerals					
			Mode	Major elements	Minor elements	Quartz	Biotite	Garnet	Plagioclase	K feldspar	Accessories
Emeryville (Group 1)	Bgn 27	Least injected	3000 counts	Chemical analyses	Spectrographic analyses	Inferred	Complete analyses	Not present	Calculated from indices of refraction	Partial analyses	Largely assumed*
	Qb 226	Incipiently granitized	3000 counts	Chemical analyses	Spectrographic analyses	Inferred	Complete analyses	Complete analyses	Calculated from indices of refraction	Analogy	Largely assumed*
	Qb 228	Least injected	3000 counts	Chemical analyses	Spectrographic analyses	Inferred	Complete analyses	Complete analyses	Calculated from indices of refraction	Partial analyses	Largely assumed*
Edwards (Group 2)	Qb 233	Incipiently granitized	4500 counts	Chemical analyses	Spectrographic analyses	Inferred	Complete analyses	Complete analyses	Calculated from indices of refraction	Analogy	Largely assumed*
	Qb 100	Incipiently granitized	4500 counts	Chemical analyses	Spectrographic analyses	Inferred	Complete analyses	Complete analyses	Calculated from indices of refraction	Partial analyses	Largely assumed*
Russell (Group 3)	Bgn 9	Incipiently granitized	4500 counts	Chemical analyses	Spectrographic analyses	Inferred	Complete analyses	Complete analyses	Calculated from indices of refraction	Analogy	Largely assumed*
	Bgn 14	Least injected	3000 counts	Chemical analyses	Spectrographic analyses	Inferred	Complete analyses	Complete analyses	Calculated from indices of refraction	Analogy	Largely assumed*
South Pierrepont (Group 4)	Bgn 25	Incipiently granitized	4500 counts	Chemical analyses	Spectrographic analyses	Inferred	Complete analyses	Complete analyses	Calculated from indices of refraction	Analogy	Largely assumed*
Colton (Group 5)	Bgn 18	Least injected	4500 counts	Chemical analyses	Spectrographic analyses	Inferred	Complete analyses	Complete analyses	Calculated from indices of refraction	Partial analysis	Largely assumed*
	Bgn 21	Least injected	4500 counts	Chemical analyses	Spectrographic analyses	Inferred	Complete analyses	Complete analyses	Calculated from indices of refraction	Partial analysis	Largely assumed*

* Ti content of magnetite and Na content of muscovite determined by chemical analysis

of the high quality necessary for sophisticated calculations of distribution factors, of the extent of ideality of solid solutions, and so on.

The available data on compositions of 10 of the most intensively studied rocks of the paragneiss are given in Table 29. In 5 rocks (Bgn 27,

TABLE 30.—COMPARISONS OF CHEMICAL COMPOSITIONS OF THE PARAGNEISS AS OBTAINED FROM DIRECT CHEMICAL ANALYSES OF THE TOTAL ROCK AND BY USE OF MODES AND CHEMICAL ANALYSES OF CONSTITUENT MINERALS

Oxide	Emeryville (Group 1)							Colton (Group 5)						
	1 Bgn 26	2 Bgn 27	3 Qb 230	4 Qb 227	5 $\bar{x}$ AR	6 $\bar{x}$ CM	7 MDM	8 Bgn 18	9 Bgn 20	10 Bgn 21	11 Bgn 19	12 $\bar{x}$ AR	13 $\bar{x}$ CM	14 MDM
SiO_2	.30	.13	.65	.20	68.93	68.42	.51	.40	.39	.35	.66	64.97	64.69	.28
TiO_2	.00	.04	.03	.02	.66	.67	.01	.12	.05	.05	.22	.64	.60	.04
Al_2O_3	.05	.40	.30	.11	14.89	14.83	.06	.23	.66	.51	.31	17.43	17.42	.01
Fe_2O_3	.04	.05	.04	.06	.77	.79	.02	.17	.11	.22	.40	.70	.51	.19
FeO	.10	.30	.15	.30	3.64	3.85	.21	.22	.18	.14	.50	5.04	5.30	.26
MnO	.02	.01	.01	.02	.06	.05	.01	.01	.00	.00	.15	2.29	2.28	.01
MgO	.10	.20	.04	.12	1.75	1.87	.12	.20	.04	.15	.60	.06	.06	.00
CaO	.10	.10	.10	.04	2.18	2.21	.03	.27	.20	.12	.30	3.68	3.90	.22
Na_2O	.06	.00	.10	.07	3.52	3.54	.02	.22	.11	.24	.12	3.57	3.52	.05
K_2O	.11	.26	.03	.17	2.89	2.80	.09	.05	.12	.23	.20	1.39	1.24	.15
H_2O	.02	.09	.04	.04	.70	.70	.00	.07	.11	.10	.09	.45	.37	.08

Columns 1–4. Mean deviations of chemical analyses of the indicated rocks, from the chemical compositions of rocks as determined by use of modes and the chemical analyses of minerals.

Column 5. Mean of the four chemical analyses of rocks posted in columns 1–4, Emeryville area (Engel and Engel, 1958a, Tables 7, 9)

Column 6. Mean of the compositions of the rocks in columns 1–4 as determined by use of modes (Engel and Engel, 1958a, Tables 7, 9) and the chemical analyses of constituent minerals

Column 7. Mean deviations of the two means posted in columns 5 and 6

Columns 8–11. Mean deviations of chemical analyses of the indicated rocks, from the chemical compositions of rocks as determined by use of modes and the chemical analyses of minerals.

Column 12. Mean of the four chemical analyses of rocks posted in columns 8–11, Colton area (Engel and Engel, 1958a, Tables 7, 9)

Column 13. Mean of the compositions of the rocks in columns 8–11 as determined by use of modes (Engel and Engel, 1958a, Table 7) and the chemical analyses of constituent minerals

Column 14. Mean deviations of the two means posted in columns 12 and 13

The data do serve as a guide and as a basis on which more refined and elegant studies may be built. There appear to be few other paragneisses of this composition and grade of metamorphism with greater potential for partitioning studies, and the stage would seem set for some very precise work in the future.

The distribution of major and trace elements among minerals in the less injected parts of the paragneiss took place under near-equilibrium conditions. This is indicated by the extremely uniform textures and simple consistent mineralogy of the paragneiss in each area and by the systematic relations between composition of the minerals, textures of rocks, and the grade of metamorphism.

Qb 228, Qb 100, Bgn 18, Bgn 21) the compositions of each constituent mineral (except plagioclase and accessories) and of the total rock have been determined by chemical and spectrochemical analyses, and the composition of the plagioclase is known from optical data. In 5 other specimens (Qb 226, Qb 233, Bgn 9, Bgn 14, Bgn 25) the compositions of all host rocks and major constituent minerals except the feldspars are known within the precision of the analyses; in these 5 rocks the composition of the feldspars is known from optical data and, in the instance of the K feldspars, by analogy with the several analyzed specimens.

Five of these rocks are least altered gneiss (Bgn 27, Qb 228, Bgn 14, Bgn 18, Bgn 21), and

TABLE 31.—FRACTIONATION OF ELEMENTS BETWEEN COEXISTING BIOTITE AND GARNET

The analyses from which the ratios are derived are given in Tables 13 and 25.

Area	Sample	$\frac{Fe^{++}+Mn}{Mg}$ B*	$\frac{Fe^{++}+Mn}{Mg}$ G†	$\frac{Ba\ B}{Ba\ G}$	$\frac{Ca\ B}{Ca\ G}$	$\frac{Co\ B}{Co\ G}$	$\frac{Cr\ B}{Cr\ G}$	$\frac{Cu\ B}{Cu\ G}$	$\frac{Fe\ B}{Fe\ G}$	$\frac{Mg\ B}{Mg\ G}$	$\frac{Mn\ B}{Mn\ G}$	$\frac{Ni\ B}{Ni\ G}$	$\frac{Sc\ B}{Sc\ G}$	$\frac{Ti\ B}{Ti\ G}$	$\frac{V\ B}{V\ G}$	$\frac{Y\ B}{Y\ G}$	$\frac{Yb\ B}{Yb\ G}$	$\frac{Zr\ B}{Zr\ G}$
Emeryville (Group 1)	Qb 226	1.84	8.96	76	.14	3.3	7.2	42	.61	2.33	.026		.10	130	16.4	.014	.0058	.47
	Qb 228	2.90	12.00	107	.10	5.0	12.5	12	.66	2.31	.016	2.2	.10	165	16.7	.030	.0087	.58
	Mean	2.37	10.48	92	.12	4.2	9.9	27	.64	2.32	.021		.10	148	16.6	.022	.007	.53
Edwards (Group 2)	Qb 233	2.18	7.46	79	.21	1.9	7.0	4.7	.58	1.72	.015	4.1	.09	117	16.6	.023	.006	.52
	Qb 100	3.42	6.11	333	.19	1.4	3.3	1.6	.74	1.26	.123	1.1	.11	209	10.0	.018	.017	.19
	Mean	2.80	6.79	206	.20	1.7	5.2	3.2	.66	1.49	.069	2.6	.10	163	13.3	.021	.011	.36
Russell (Group 3)	Bgn 9	2.07	7.87	35	.55	2.0	3.8	3.3	.51	1.89	.034	20.0	.10	157	16.1	.042	.015	.67
	Bgn 14	2.81	7.68	120	.17	2.9	4.4	14.0	.65	1.67	.023	20.0	.08	103	16.7	.012	.0066	.85
	Mean	2.44	7.78	78	.36	2.5	4.1	8.7	.58	1.78	.029	20.0	.09	130	16.4	.027	.011	.76
South Pierrepont (Group 4)	Bgn 25	1.76	6.08		.04				.59	1.91				118				
Colton (Group 5)	Bgn 18	1.57	5.05	500	.74	1.7	4.5	50.0	.50	1.46	.018	20.0	.08	199	14.3	.055	.083	1.33
	Bgn 21	1.77	4.91	1100	.47	1.7	4.4	2.5	.56	1.45	.018	20.0	.07	115	15.0	.050	.057	.67
	Mean	1.67	4.98	800	.61	1.7	4.5	26.2	.53	1.45	.018	20.0	.08	157	14.7	.053	.070	1.00

* B = Biotite

† G = Garnet

5 are incipiently granitized gneiss (Qb 226, Qb 233, Qb 100, Bgn 9, Bgn 25). Both types are distributed from Emeryville to Colton (Table 29; Pl. 1) so that the partitioning of elements

TABLE 32.—RATIO Fe^{++} + Mn:Mg IN COEXISTING GARNET, BIOTITE, AND HOST ROCKS,* AND THE RATIO $\frac{Fe^{++} + Mn:Mg\ Garnet}{Fe^{++} + Mn:Mg\ Biotite}$

Area and group	Number of samples	Fe^{++}Mn:Mg			(Fe^{++} + Mn: Mg Garnet) / (Fe^{++} + Mn: Mg Biotite)
		Garnet	Biotite	Rock	
Edwards (Group 2)..	2	6.79	2.80	3.23	2.43
Russell (Group 3)..	2	7.78	2.44	2.56	3.19
South Pierrepont (Group 4)..	1	6.08	1.76	2.19	3.45
Colton (Group 5)..	2	4.98	1.67	2.93†	2.98

* Source data for rocks from Engel and Engel (1958a, Tables 7, 9).

† FeO value for rock Bgn 21, in Engel and Engel (1958a, Table 7) should read 3.86 instead of 0.86.

may be plotted as a function of both grade of metamorphism and the change in bulk composition of the host rock.

The partitioning of elements between biotite and garnet is detailed in Table 31. These data are fairly complete for the 9 rocks shown and are discussed in the succeeding pages. Where possible they are related to the partitioning of elements in associated minerals and to the variations in composition of the total rock.

Iron-Magnesium-Manganese.—The important mafic minerals in the paragneiss are biotite and garnet. Magnetite, ilmenite, and minute traces of sulfides (largely pyrite) are present in most specimens, but these minerals commonly total less than 0.3 volume per cent of the rock and average about 0.15 per cent (Part I, Engel and Engel, 1958a, Tables 7, 9). Their presence suggests that the system was saturated in Fe and Ti, yet the amounts of these and chemically coherent elements tied up in the opaque minerals are quite small. The garnet characteristically is more ferrophile than biotite, and the ratios of Fe^{++} + Mn:Mg in coexisting garnet and biotite decrease by about 50 per cent with increasing grade of metamorphism (Tables 31, 32; Fig. 11).

The ratios of Fe_2O_3:FeO in both biotite and coexisting garnet decrease with increasing grade of metamorphism, although at different rates (Fig. 12). The most profound change in Fe_2O_3:FeO in the garnets, which is not indicated in Figure 11, is from the spessartitic almandites in the injected gneiss at Emeryville (Fe_2O_3:FeO = 0.07) to the Mn-poor almandites to the northeast (Fe_2O_3:FeO = 0.03). The ratio Fe_2O_3:FeO in biotite to Fe_2O_3:FeO in garnet commonly ranges from 1 to 3.

Titanium.—The abundance of TiO_2 in the paragneiss permits its fractionation to be followed in some detail. Of the major constituents, biotite is most enriched in TiO_2; it contains from 130 to more than 200 times more than coexisting garnet (Table 31). The garnet has in turn about 5 to 10 times as much TiO_2 as coexisting K feldspar, depending upon grade of metamorphism, but the ratio of Ti in garnets to Ti in plagioclase is close to 1 throughout the gneiss. Although the TiO_2 in both biotite and garnet increases by about 75 per cent between Emeryville and Colton the TiO_2 content of the host rock remains essentially constant. This is possible because of the associated decrease in biotite (Fig. 1).

Potassium and sodium.—The Na and K in most of the paragneiss[16] under discussion are distributed among biotite, plagioclase, and K feldspar. In least altered gneiss nearly all the K is in the biotite, and the Na is in the plagioclase. With increasing grade of metamorphism, the decrease in biotite and increase in total plagioclase (and in its An content) are accompanied by a decrease in the K_2O:Na_2O ratio of the total rock (from 0.7 to 0.4). Actually the amount of Na_2O in the rock remains about the same, whereas large amounts of K_2O are expelled (Fig. 1).

In the variously granitized specimens of gneiss, appreciable K and some Na are locked in the K feldspar. The distribution of Na and K between the plagioclase and K feldspar in 11 rocks is posted in Table 1. Experimental studies of the feldspars, especially ternary systems of Or–Ab–An (Tuttle and Bowen, 1958; Yoder, Stewart and Smith, 1957), indicate the possible value of these distribution data, especially as guides to *T* and *P* during crystal growth. In Figure 12, the orientations of the tie lines con-

[16] Parts of least altered and granitized gneiss at Emeryville contain up to 2 volume per cent muscovite, and traces of sericite appear in many plagioclases throughout the gneiss (Table 2).

necting 5 (typical ?) coexisting feldspars in the gneiss (Table 1) are plotted on the ternary diagram Or–Ab–An.

As noted in the preceding discussion of K feldspar, the plagioclase and the analyzed K feldspar are in semiseparate folia, and possibly the two minerals did not crystallize in equilibrium. At least slight departure from equilibrium is also suggested by the small deviations in trend between some of the tie lines in Figure 12 and those determined experimentally by Yoder, Stewart, and Smith (1957, p. 212). The largest deviations are in the tie lines of the feldspar pairs Bgn 18 and Bgn 21 (Colton). Specifically, the An content of the plagioclase appears slightly high relative to the Na content of the K feldspar. This is the sort of discordance expectable if the K feldspar were crystallized from a more felsic fluid either mobilized from, or injected into, the gneiss.

Scandium, Yttrium, and Ytterbium.—The elements Sc, Y, Yb are concentrated in the garnets, presumably in association with Fe^{++}, Mg, and Mn. Biotite is the only other major constituent with detectable amounts of Sc, Y, and Yb. The fractionation ratio of Sc in biotite to Sc in garnet is about constant at 0.1 to 0.08. Y in biotite to Y in garnet ranges from .02 at Emeryville to .05 at Colton. Yb in biotite to Yb in garnet ranges from .007 at Emeryville to .07 at Colton (Table 31; Fig. 11). In the least altered gneiss between Emeryville and Colton Fe^{++} and Mg increase, but there is little or no change in the concentrations of Mn, Sc, Y, and Yb; the concentrations of these elements in the rock are in parts per million: Mn, 345; Y, 50; Sc, 15; Yb, 3 (Part I, Engel and Engel, 1958a, Table 8).

Cobalt, Chromium, Nickel, and Vanadium.—Co, Cr, Ni, and V are more concentrated in the biotites than in the garnets (Table 31). Vanadium exhibits the most asymmetrical fractionation between biotite and garnet with 15 to 16

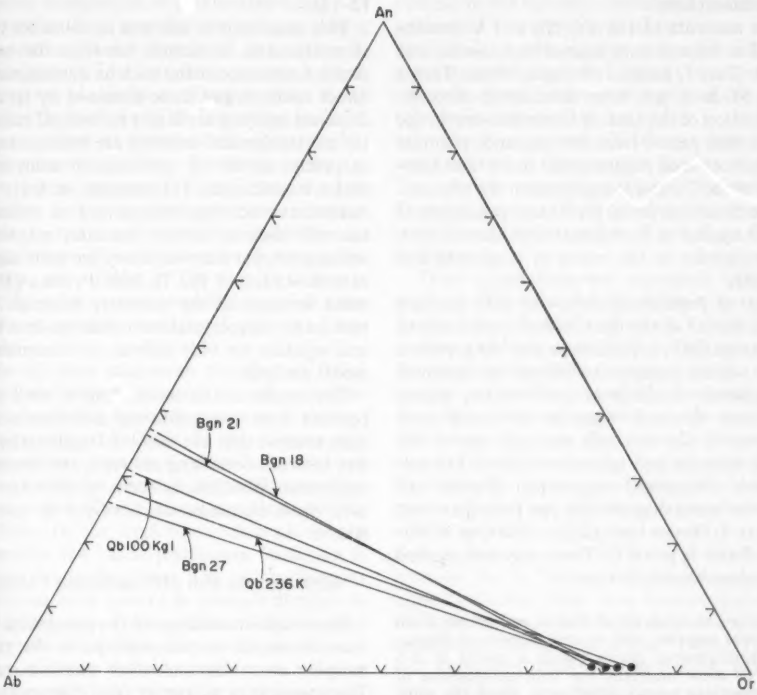

FIGURE 12.—TERNARY PLOT OF THE ORIENTATIONS OF TIE LINES BETWEEN FIVE PAIRS OF COEXISTING FELDSPARS IN THE GNEISS COMPLEX
The compositions of these feldspars are given in Table 1.

times more V in the biotite. Fractionation ratios between biotite and garnet from the other elements are: Cr = 4–10; Ni = 2–20; Co = 2–4. All these ratios except that for Ni decrease northeast of Emeryville, with increasing grade of metamorphism.

The amounts of Co, Cr, Ni, and V concentrated in the accessory magnetite, ilmenite, and pyrite (Part I, Engel and Engel, 1958a, Tables 7, 8, 9) have not been determined directly. Subtraction of the total of these elements in the biotite and garnet from the amounts recorded for the host rock suggests that more than four-fifths of the Co and V is present in the silicates. About three-fourths of the Ni and two-thirds of the Cr appear to be segregated in the silicates, the remainder in the traces of magnetite and ilmenite.

Tests of partitioning data.—As a by-product of the studies of the distribution of elements in the paragneiss it is possible to test the precision of the various analyses as follows: the chemical composition of the host rock—major, minor, and trace elements—may be calculated from analyses of the minerals through use of the modal analyses and mineral densities. The calculation of chemical composition of each rock from the mineralogical data has been described in Part I (Engel and Engel, 1958a, p. 1380–1382; Table 5, p. 1381). The conclusion reached for major elements is that

> "... given an analysis of biotite and garnet from the gneiss together with a careful mode of reasonably homogeneous gneiss within a radius of 2–3 miles of these minerals, the bulk composition of the gneiss can be calculated with about the same precision as the 34 analysts obtained in their work in the granite G-1 (Fairbairn *et al.*, 1951; Fairbairn, 1953). In areas of this size involving the most homogeneous parts of the gneiss where the compositions of more than one biotite and one garnet are known, the precision of such calculations of the concentration of most major elements in the gneiss is greater than that achieved by the 34 analysts of G-1" (p. 1382).

The ideal relations are obviously those in which the chemical compositions of minerals are available for the most nearly homogeneous hand specimen. The potentials of this approach are illustrated by the data in Table 30, which indicate the specific differences in chemical compositions of some of the rocks as determined by (1) direct chemical analysis and (2) use of detailed modes and the chemical compositions and densities of constituent minerals. For those specimens with more inhomogeneous textures the results are less precise.

The mean compositions of the four specimens in each group, as determined from modes and from the compositions of constituent minerals, are essentially identical with the mean compositions of these rocks as determined by direct chemical analysis (Table 30, columns 5–6 and 12–13).

This conclusion is not true however for trace elements. The deviations between the trace-element contents of the rock as determined by direct analyses and those obtained by spectrochemical analyses are larger for several reasons: (1) spectrochemical analyses are not as accurate or precise as are the gravimetric analyses of major constituents; (2) contents of trace elements have not been determined in accessory minerals such as zircon, ilmenite, magnetite, and apatite, yet they obviously are quite significant in studies of Zr, Ti, Mn, P, etc.; (3) the exact amounts of the accessory minerals in a rock or as microcrystalline inclusions in a mineral separate are very difficult to determine by modal analysis.

The results of the calculations of rock composition from composition of constituent minerals suggest that the derived fractionation ratios between coexisting minerals are internally consistent. Whether, in terms of ultimate values, these numbers are accurate is another matter.

Mineralogy and Metamorphic Facies

The simple mineralogy of the paragneiss may leave uncertain its relationships to the metamorphic assemblages of other classic terranes. The appearance of garnet (and disappearance of muscovite) at Edwards constitutes the single classical "isograd" in a belt of paragneiss almost 75 miles long (Philadelphia, New York, to Colton, New York). There is no isograd of the classical type from Edwards to Colton, a distance along the strike of almost 30 miles, yet the temperature differences involved seem to be 100°C, and the variations in many properties of the mineral species are systematic and large. The mineralogical data are far more meaningful if presented in the context of the mineralogy of associated metasedimentary rocks, especially the amphibolites, siliceous magnesian marble, and pelites. These relationships are summarized in Table 33 and discussed briefly below.

Siliceous magnesian marbles lie on opposite sides of the paragneiss throughout the Emeryville-Colton region, and amphibolite layers an inch to 200 feet thick are intercalated in it (A. Engel, 1956; Engel and Engel, 1953; 1958a; 1958b). Just south of the paragneiss at Emery-

ville is a typical metapelite (Brown and Engel, 1956). The mineral assemblages in these diverse units should ring familiar bells for everyone.

The amphibolites that form thin but persistent interlayers in the paragneiss have the bulk chemical composition of a saturated basalt. At Emeryville these amphibolites average quartz 10, andesine 20, hornblende 67, magnetite, ilmenite, and sulfides 2 (volume per cent). Some have traces of pale-green clinopyroxene. Many are slightly biotitic. With increasing grade of metamorphism (at Edwards) clinopyroxene becomes common. At Russell orthopyroxene appears in most specimens. Near Colton the amphibolite layers in the paragneiss average quartz 1, calcic andesine 39, hornblende 22, clinopyroxene 22, orthopyroxene 10, magnetite, ilmenite, and sulfides 4. The rock at Colton is actually a pyroxene amphibolite, and some interlayers contain less than 5 per cent hornblende.

The siliceous magnesian marbles on either side of the paragneiss contain various assemblages depending upon the presence or absence of magnesium and alkalis. In the siliceous, calcitic marbles, quartz and calcite have coexisted without reaction throughout the Emeryville–Colton region (Part I, Engel and Engel, 1958a, p. 1386). In the argillaceous dolomites both phlogopite and calcic plagioclase (andesine to anorthite) are ubiquitous. The quartz-dolomite interlayers have reacted to produce diopside as the highest-rank metamorphic silicate at both Emeryville and Colton. In general, the diopside in the siliceous dolomites occurs as reaction rims or coronas separating quartz and dolomite. These rims tend to be thickest in the Colton region, and in many beds the least abundant primary constituent—either quartz or dolomite—is used up in the metamorphic reaction that formed the diopside. This means more CO_2 was evolved from the marbles at Colton than at Emeryville. The relations indicate that the evolution of CO_2 was dependent upon the rate of either (1) loss of CO_2 from the rock system at any point or (2) diffusion of Ca, Mg, and Si through the initially formed diopside sheath that separated the dolomite from quartz. The higher temperatures at Colton could have facilitated diffusion, but the decrease in H_2O in the system may have retarded the process.

The pelitic unit is found in association with the paragneiss at and near Emeryville (a member of unit No. PS2 on Plate 1, *in* Brown and Engel, 1956). It is a metamorphosed K-rich shale consisting of the assemblage quartz, biotite, sillimanite, and muscovite.

If we follow classical terminology, the metasedimentary sequence at Emeryville is typical of the upper part of the amphibolite facies (Table 33). The transition to granulite facies occurs at Edwards, with the formation of K feldspar at the expense of muscovite in the paragneiss and the appearance of pyroxenes in the amphibolite at the expense of hornblende. The sequence northeast of Edwards represents a transition from the lowermost part of the granulite facies, almost (but not quite) to the typical, anhydrous pyroxene-K feldspar-garnet-rich manifestation of the granulite facies at Colton. The mineral assemblages in the Emeryville–Colton region may be related to the principal metamorphic facies, and to T and P, as shown in Figure 2.

These correlations are consistent with the work of Eskola and with interpretations by Barth, Tilley, Turner, and others. Turner (Fyfe, Turner, and Verhoogen, 1958, p. 166) notes that the amphibolite facies is marked by the presence of amphiboles (in amphibolites) and mica (in pelites and metagraywackes). He concludes that almandite is a characteristic mineral if associated with muscovite. Typical minerals of the granulite facies are quartz, K feldspar, garnet, plagioclase, sillimanite, hypersthene, diopside, and calcite. Tilley notes (oral communication) that the granulite facies is ideally anhydrous. Barth, Turner, and Eskola agree, but typical granulite facies "may be complicated by the appearance of additional minerals, notably hornblende, biotite and zoisite" (Fyfe, Turner and Verhoogen, 1958; *see also* Eskola, 1957, p. 106). Actually the progressive metamorphism preserved between Emeryville and Colton is a superb example of transition from typical hydrous amphibole-biotite-rich amphibolite facies to only slightly hydrous garnet-plagioclase-K feldspar-pyroxene-rich (lower) granulite facies. The progressive, albeit incomplete, dehydration and decarbonation of the systems are systematic and clearly evident. Only a small additional rise in T (and P?) and loss of water at Colton would seem to have been necessary to produce the classical granulite facies. In the range of composition of specimens that constitute the average rocks at Colton are garnet-plagioclase paragneiss with less than 6 per cent biotite[17] and two-pyroxene-plagioclase gneisses ("amphibolites") with less than 2 per cent hornblende.

Because of the rarity of rocks transitional be-

[17] We searched carefully for pyroxene-garnet-plagioclase paragneiss but have yet to find a specimen, although these exist locally in the massif.

tween amphibolite and granulite facies Fyfe, Turner, and Verhoogen (1958, p. 233) only

"tentatively correlate the pyroxene granulite subfacies with higher temperatures than those of the almandine amphibolite facies".

ical and chemical properties we have investigated, the variations in composition of mineral phases, especially biotites and garnet, seem ideal. Lines of equal "grade" of metamorphism can readily be drawn upon increments of change

TABLE 33.—METAMORPHIC FEATURES AND MINERAL ASSEMBLAGES IN THE EMERYVILLE–COLTON REGION, NORTHWEST ADIRONDACK MOUNTAINS, NEW YORK

FEATURE	EMERYVILLE ← 35 miles → COLTON
Premetamorphic sediment	← little lateral variation →
Temperature of metamorphism	500° C — uniform increase → 600° C
Depth of metamorphism	5 miles ? — increase → 7 miles ?
Postmetamorphic texture	← increase in grain size × 5 → more inequigranular
Metamorphic facies	Upper amphibolite → lower granulite
Changes in bulk composition due to metamorphism	← decrease in H_2O, CO_2, Si, K → increase in Fe, Ca, Mg

ROCK TYPE	VOLUME PER CENT*	ASSEMBLAGE
Paragneiss	30	plagioclase, quartz, biotite, muscovite → plagioclase, quartz, biotite, garnet
Amphibolite	10	hornblende, plagioclase, quartz → plagioclase, hornblende, clino- and orthopyroxene
Siliceous ls.	10	quartz, calcite → quartz, calcite
Siliceous dol.	50	dolomite, quartz, diopside → dolomite, diopside, quartz

* Includes about half the metasedimentary rock section exposed in northwest Adirondack Mountains

In the northwest Adirondacks the transition between amphibolite facies and the so-called hornblende (or biotite) granulite subfacies has evolved along gradients in temperature (and water pressure?) during one metamorphism. The region also appears to represent a type of ultrametamorphism, or selective fusion (solution) and migration of the more mobile felsic constituents.

The simplicity of the mineralogy within the paragneiss points up an inherent value in studies of the physical and chemical properties of constituent minerals. One isograd in 35 to 40 miles is sparse delineation of the beautifully systematic progressive metamorphism of a clastic metasediment. But any of several properties of constituent minerals may be employed to complement the conventional isograds, as an indicator of a rise in "metamorphic grade." The consistent changes in color of biotite alone make possible the use of a "colorgrad." Of the physical and chemical properties we have investigated, the variations in composition of mineral phases, especially biotites and garnet, seem ideal. Lines of equal "grade" of metamorphism can readily be drawn upon increments of change of such elements as Fe, Mg, Mn, or Ti in biotite and garnet. Some of these elements or their oxides are easily determined by one of several well-established techniques. Concentrations of Ti and Mn, which in biotites and garnets appear to be highly sensitive to grade of metamorphism, are easily measured by emission spectroscopy, rapid methods of chemical analysis, or X-ray fluorescence. Miyashiro (1953) has emphasized the value of Mn in almandites as a possible guide to grade of metamorphism.

Perhaps even better compositional indicators of the grade of metamorphism in the paragneiss and other units having a given mineral assemblage are ratios of two elements in constituent minerals, especially where one element increases and the other decreases. The ratios of Ti:Mn, Mn:Fe, or Fe:Mg in biotites and in garnets are examples.

One great value of the use of such ratios to define traces of planes of equal grade of meta-

morphism is that they offer increments of change appreciably larger than the change in concentration of any one element or oxide in the minerals. The increase in absolute amount of TiO_2 in biotite between Emeryville and Colton, for example, is about 2 weight per cent (from 3 to 5). The decrease in Mn in biotite is from 1400 ppm at Emeryville to 90 ppm at Colton; the change in the ratio TiO_2:MnO in biotite between Emeryville and Colton is by a factor of 32, however, from 16 to more than 500. The ratios Mn:Ti, Mn:Fe, or Fe:Mg in garnets have analogous advantages over the change in concentration of a single element or oxide. The ratio of TiO_2:MnO in garnets changes from .003 at Emeryville to 0.06 at Colton. Quite detailed lines of equal intensity or "grade" of metamorphism may be defined upon larger-scale maps of the Emeryville-Colton region via this technique.

Ratios of major elements, such as Fe:Mg or Fe^{++}+Mn:Mg, are of equal or greater value but are more difficult to obtain by analysis. The number and variety of these ratios available for definition of metamorphic grade, of equilibrium relations and related problems are very large (Tables 13, 25; Figs. 6, 8). Eventually the relationships of the most meaningful ratios to *T* and *P* may be established as a broad and powerful amplification of any petrogenic grid (Bowen, 1940).

The use of chemical compositions of minerals as a guide to grade of metamorphism is but a single facet of the obvious fact that metamorphic processes will not be understood until we know the physical and chemical properties of metamorphic rocks and their constituent minerals.

References Cited

Ambrose, J. W., 1936, Progressive kinetic-metamorphism of the Missi series near Flinflon, Manitoba: Am. Jour. Sci., v. 32, p. 257–286

Barth, T. F. W., 1936, Structural and petrologic studies in Dutchess County, New York, Pt. II: Geol. Soc. America Bull., v. 47, p. 775–850

—— 1956, Studies in gneiss and granite: Skrifter Norske Vidensk Akad, Mat.-Naturv. Klasse, no. 1, p. 1–35

Billings, M. P., 1937, Regional metamorphism of the Littleton-Moosilauke area, New Hampshire: Geol. Soc. America Bull., v. 48, p. 463–566

Bowen, N. L., 1940, Progressive metamorphism of siliceous limestone and dolomite: Jour. Geology, v. 48, p. 225–274

Brown, J. S., and Engel, A. E. J., 1956, Revision of Grenville stratigraphy and structure in the Balmat-Edwards district, northwest Adirondacks, New York: Geol. Soc. America Bull., v. 67, p. 1599–1622

Crowley, M. S., and Roy, Rustum, 1958, Possible Al-Si ordering in sheet structures (Abstract): Geol. Soc. America Bull., v. 69, p. 1549–1550

DeVore, G. W., 1953, Temperature and the distribution of trace elements (Abstract): Geol. Soc. America Bull., v. 64, p. 1412

—— 1955a, The role of adsorption in the fractionation and distribution of elements: Jour. Geology, v. 53, p. 159–190

—— 1955b, Crystal growth and the distribution of elements: Jour. Geology, v. 63, p. 471–494

Engel, A. E. J., 1956, Apropos the Grenville, p. 74–98 *in* Thomson, J. E., *Editor*, The Grenville problem: Royal Soc. Canada Special Pub. no. 1, 119 p.

Engel, A. E. J., and Engel, Celeste G., 1953, Grenville series in the northwest Adirondack Mountains, New York. Part II: Origin and metamorphism of the major paragneiss: Geol. Soc. America Bull., v. 64, p. 1049–1097

—— 1958a, Progressive metamorphism and granitization of the major paragneiss, northwest Adirondack Mountains, New York, Part I, Total rock: Geol. Soc. America Bull., v. 69, p. 1369–1414

Engel, Celeste G., and Engel, A. E. J., 1958b, Progressive metamorphism of amphibolite, northwest Adirondack Mountains, New York (Abstract): Geol. Soc. America Bull., v. 69, p. 1559

Engelhardt, Wolf von, 1936, Die Geochemie des Barium: Chem. Erde, v. 10, 246 p.

Eskola, P., 1952, On the granulites of Lapland: Am. Jour. Sci., Bowen Volume, pt. I, p. 133–171

—— 1957, On the mineral facies of charnockites: Jour. Madras Univ., v. 27, p. 101–119

Eugster, H. P., 1955, Experimental determination of the distribution factor of trace elements (Abstract): Geol. Soc. Am. Bull., f. 66, p. 1555

Fairbairn, H. W., 1953, Precision and accuracy of chemical analyses of silicate rocks: Geochim. et Cosmoschim. Acta, v. 4, p. 143–156

Fairbairn, H. W., and others, 1951, A cooperative investigation of precision and accuracy in chemical, spectrochemical, and modal analysis of silicate rocks: U. S. Geol. Survey Bull., 980, 54 p.

Fleischer, Michael, 1937, The relation between chemical composition and physical properties in the garnet group: Am. Mineralogist, v. 22, p. 751–759

Ford, W. E., 1915, A study of the relations existing between the chemical, optical and other physical properties of the members of the garnet group: Am. Jour. Sci., v. 40, p. 33–49

Fyfe, W. S., Turner, F. J., and Verhoogen, Jean, 1958, Metamorphic reactions and metamorphic facies: Geol. Soc. America Mem. 73, 259 p.

Goldschmidt, V. M., 1921, Die Injektionsmetamorphose im Sravanger-Gebiete: Oslo Vidensk. Skrift. I Mat. Naturv. Klasse 1920, no. 10, 142 p.

—— 1954, Geochemistry: Oxford, Clarendon Press, 730 p.

Goldsmith, J. R., and Laves, F., 1954a, The microcline-sanadine stability relations: Geochim. et Cosmochim. Acta, v. 5, p. 1–19

—— 1954b, Potassium feldspars structurally intermediate between microcline and sanidine: Geochim. et Cosmochim. Acta, v. 6, p. 100–118

Gower, J. A., 1957, X-ray measurement of the

iron-magnesium ratio in biotites: Am. Jour. Sci., v. 255, p. 142–156
Hall, A. J., 1941, The relation between chemical composition and refractive index in the biotites: Am. Mineralogist, v. 26, p. 34–41
Harker, Alfred, 1932, Metamorphism: London, Methuen, 360 pages
Heier, K. S., 1957, Phase relations of potash feldspar in metamorphism: Jour. Geology, v. 65, p. 468–479
Heier, K. S., and Taylor, S. R., 1959, Distribution of Li, Na, K, Rb, Cs, Pb, and Tc in southern Norwegian pre-Cambrian alkali feldspars: Geochim. et Cosmoschim. Acta, v. 15, p. 284–304
Heinrich, E. Wm., 1946, Studies in the mica group: the biotite-phlogopite series: Am. Jour. Science, v. 244, p. 836–844
Hendricks, S. B., and Jefferson, M. E., 1939, Polymorphism of the micas: Am. Mineralogist, v. 24, p. 729–771
Hurst, C. J., 1957, Polymorphism of micas in the Mineral Bluff and Epworth quadrangles, Georgia: Geol. Soc. America Bull., v. 68, p. 1581–1584
Ingerson, Earl, 1955, Methods and problems of geologic thermometry: Econ. Geology, Fiftieth Anniversary Volume, p. 341–410
Jaffe, H. W., 1956, Application of the rule of Gladstone and Dale to minerals: Am. Mineralogist, v. 41, p. 757–777
James, H. L., 1955, Zones of regional metamorphism in the Precambrian of northern Michigan: Geol. Soc. America Bull., v. 66, p. 1455–1488
Kunitz, Wilhelm, 1936, Beitray zur Kenntnis der magmatischen Assoziationen III. Die Rolle des Titans and Zirkoniums in den gesteinsbildenden Silikaten. Neues Jahrb. Mineral. Geol., Beilage-Bd. 70, A, 385 p.
Larsen, E. S., and Berman, Harry, 1934, Microscopic determinations of the nonopaque minerals: U. S. Geol. Survey Bull. 848, 266 p.
Lee, D. E., 1958, A chlorine-rich biotite from Lemhi County, Idaho: Am. Mineralogist, v. 53, p. 107–111
Levinson, A. A., and Heinrich, E. Wm., 1954, Studies in the mica group: Single crystal data on phlogopites, biotites and manganophyllites: Am. Mineralogist, v. 39, p. 937–945
Lundegårdh, Per H., 1947, Some aspects to the determination and distribution of zinc: Roy. Agr. Coll. Sweden Ann., v. 15, p. 1–16
Miyashiro, Akiho, 1953, Calcium-poor garnet in relation to metamorphism: Geochim. et Cosmoschim. Acta, v. 4, p. 179–208
—— 1958, Regional metamorphism of the Gosaisyo-Takanuki district in the central Abukuma Plateau: University Tokyo, Jour. Faculty Sci., sec. 2, v. 11, pt. 2, p. 219–272
Mood, A. M., 1950, Introduction to the theory of statistics: New York McGraw-Hill Book Co., 325 p.
Nockolds, S. R., 1947, The relations between chemical composition and paragenesis in the biotite micas of igneous rocks: Am. Jour. Sci., v. 245, p. 401–420
Nockolds, S. R., and Mitchell, R. L., 1948, The geochemistry of some Caledonian plutonic rocks: A study in the relationship between the major and trace elements of igneous rocks and their minerals: Royal Soc. Edinburgh Trans. v. 61, pt. 2, p. 533–575
Oftedal, Ivar, 1943, Scandium in biotite as a geologic thermometer: Norsk Geologisk Tidsskrift, v. 23, p. 202–213
Perrin, R., 1956, Granite again: Am. Jour. Sci., v. 254, p. 1–18.
Pettijohn, F. J., 1957, Sedimentary rocks: New York, Harper and Bros., 690 p.
Phillips, F. C., 1930, Some mineralogical and chemical changes induced by progressive metamorphism: Min. Mag., v. 22, p. 239–256
Raguin, E., 1946, Geologie du granite: Masson et Cie, Paris 158 p.
Ramberg, Hans, 1949, The facies classification of rocks: a clue to the origin of quartzofeldspathic massifs and veins: Jour. Geology, v. 57, p. 18–54
—— 1952, The origin of metamorphic and metasomatic rocks: Chicago, Chicago Univ. Press, 317 p.
Rankama, Kalervo, and Sahama, Th. G., 1950, Geochemistry: Chicago, Univ. Chicago Press, 912 p.
Read, H. H., 1948, Granites and granites, p. 1–19, *in* Origin of granite, Gilluly, James, Chairman: Geol. Soc. America Mem. 28, 139 p.
Sandell, E. B., and Goldich, S. S., 1943, The rarer metallic constituents of some American igneous rocks, Pt. 1: Jour. Geology, v. 51, p. 99–115
Shapiro, Leonard, and Brannock, W. W., 1956, Rapid analysis of silicate rocks: U. S. Geol. Survey Bull. 1036–C, p. 19–56
Skinner, B. J., 1956, Physical properties of end-members of the garnet group: Am. Mineralogist, v. 41, p. 428–436
Smith, J. R., and Yoder, H. S., 1956a, Variations in X-ray powder diffraction patterns of plagioclase feldspars: Am. Mineralogist, v. 41, p. 632–647
Stewart, F. H., 1950, Note on garnet crystals from Cairnie, Aberdeenshire: Min. Mag., v. 29, p. 252–253
Stevens, R. E., 1946, A system for calculating analyses of micas and related minerals to end members: U. S. Geol. Survey Bull. 950, p. 101–119
Sugi, Kenichi, 1935, A preliminary study of the metamorphic rocks of southern Abukuma Plateau: Jap. Jour. Geology and Geography, v. 12, p. 115–151
Taylor, S. R., and Heier, K. S., 1958, Alkali elements in potash feldspar from the pre-Cambrian of Southern Norway: Geochim. et Cosmoschim. Acta, v. 13, p. 293–302
Tilley, C. E., 1925, Metamorphic zones in the southern Highlands of Scotland: Geol. Soc. London Quart. Jour., v. 81, p. 100–112
Turner, F. J., 1933, The genesis of oligoclase in certain schists: Geol. Mag., v. 70, p. 529–541
Tuttle, O. F., and Bowen, N. L., 1958, Origin of granite in the light of experimental studies in the system $NaAlSi_2O_8$–$KAlSi_3O_8$–SiO_2–H_2O: Geol. Soc. America, Memoir 75, 154 p.
Weiss, Judith, 1949, Wissahickon schist at Philadelphia, Pennsylvania: Geol. Soc. America Bull., v. 60, p. 1689–1726
Wright, W. I., 1938, The composition and occurrence of garnets: Am. Mineralogist, v. 23, p. 436–449

Yoder, H. S., 1955, Role of water in metamorphism, p. 505–524 *in* Poldervaart, Arie, *Editor*, Crust of the Earth: Geol. Soc. America Special Paper 62, 762 p.

Yoder, H. S., and Eugster, H. P., 1954, Phlogopite synthesis and stability range: Geochim. et Cosmoschim. Acta, v. 6, p. 157–185

Yoder, H. S., Stewart, D. B., and Smith, J. R., 1957, Feldspars, p. 206–214 *in* Annual Report to the Director; Geophysical Lab., Carnegie Inst. Washington

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KNICKPOINT BEHAVIOR IN NONCOHESIVE MATERIAL: A LABORATORY STUDY

BY LUCIEN M. BRUSH, JR., AND M. GORDON WOLMAN

ABSTRACT

Short oversteepened reaches were molded in the beds of model channels formed in well-sorted, noncohesive sands 0.67 mm and 2 mm in diameter. In each run of the experiment a fall of 0.1 foot in a length of 1.0 foot was provided. The over-all slopes of the channels upstream and downstream from the oversteepened reach were made equivalent and ranged from 0.0012 to 0.0088. The abrupt break in the profile at the head of the oversteepened segment constituted a knickpoint. Progressive changes in the position of the knickpoint and in the slope of the oversteepened reach were measured during runs in which discharge, over-all slope, and particle size were varied independently.

In every run, the slope of the water surface and the slope of the bed below the knickpoint decreased with time. As the knickpoint moved upstream, the channel directly above the knickpoint first steepened and narrowed. Following the initial steepening, the slope became progressively less. At the lower end of the oversteepened reach, sediment eroded from above was deposited as a dune, which advanced downstream and caused the channel to widen and locally to steepen. Following the passage of the dune, the slope again flattened.

For runs with identical initial (over-all) slopes, discharge, and widths, the slope below the knickpoint decreased faster in the channel of finer sand.

The rate of change of slope in the oversteepened reach below the knickpoint depends upon the magnitude and the rate of change of erosion along this reach which, in turn, depends upon the magnitude and the rate of change of the sediment transport along the reach. If the rate of transport is great, the oversteepened slope below the knickpoint is reduced rapidly. Analysis of the data indicates that the higher the ratio of the oversteepened slope to the average slope the more rapid the rate of decrease of the oversteepened slope. These results are comparable to changes observed in natural stream channels following meander cutoffs.

The laboratory experiment confirms the observation that upstream migration of knickpoints accompanied by undiminished slopes does not occur in noncohesive, homogeneous bed material. Several hypothetical cases are discussed in which it is assumed that resistant material is present in the channel profile.

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Introduction and Acknowledgments

Knickpoints, or points along longitudinal profiles of streams at which the slope increases abruptly, have long been of interest to geologists. Numerous field observations have been made of the position, number, and geologic environment of knickpoints. Rational theories have been proposed to explain the origin, rate of movement, and method of migration of knickpoints. However, few experimental studies have been made.

TERMINOLOGY: Abrupt breaks in slope in river profiles are called by many names. All are similar in form but have different meanings to different authors. Two spellings and definitions are given in the recent American Geological Institute Glossary (1957):

"*Nickpoint*—The point of interruption of a stream profile at the head of a second-cycle valley according to the Treppen concept.

"*Knickpoint*—Points of abrupt change in the longitudinal profile of stream valleys."

Von Engeln (1949, p. 266) summarizes the concept of *Knickpunkte* or nickpoints by stating that there are two schools of thought:

"According to the precepts of the American school, knickpunkte, in the profiles of streams and in slopes, are merely the expression of an ungraded condition. As such they may be present because of marked differences in resistance to degradational attack of rock beds or structure, or are evidence of intermittent uplift. Cap-rock waterfalls, cuestas, two-cycle valleys, and fault scarps are examples.

"The Walther Penck doctrines hold that these explanations are inadequate. The examples cited are, in accordance with his approaches, only special cases, of a general phenomenon, whose occurrence need not be conditioned by such exceptional circumstances of structure and uplift. Penck contended, instead, that knickpunkte are a normal feature in the profiles of streams that descend the slopes of a dome of homogeneous rock which is experiencing continuously accelerated, but uninterrupted, uplift, and is expanding in area. Further, the knickpunkte are matched in the interstream areas by the steep fronts of piedmont treppen (piedmont bench lands)."

Knickpunkte and knickpoint (a literal English translation of the German *Knickpunkte*) are both associated with the Treppen concept which, contrary to Penck, has not been established as the general case. The American Geological Institute definition of knickpoint seems more general in that it applies to any abrupt change in the profile regardless of the cause of its initiation. In this discussion, a point of abrupt increase in slope will be called a knickpoint, because the study concerns the process of knickpoint reduction and migration and not the genesis of knickpoints under restricted or specified conditions.

A flume built by the U. S. Geological Survey at the University of Maryland, College Park, Maryland, was used for this study. The authors wish to thank Prof. John P. Miller of Harvard University, Prof. A. N. Strahler of Columbia University, and J. Hoover Mackin of the University of Washington for critically reviewing the manuscript.

Theory and Procedure

Theory of Changes Caused by Knickpoints

It is logical to suppose that knickpoint changes depend on the amount of sediment transported at any particular point in a stream. The volume of flow, channel shape, slope of the oversteepened reach, and average slope of the river are also important, but directly or indirectly all these factors contribute to the determination of the transport of the sediment load. The term oversteepened reach in this paper refers to the reach immediately downstream from the knickpoint where the slope is steeper than that in adjacent reaches. A knickpoint in a channel in which no material is being transported will remain intact indefinitely. In a stream flowing over a knickpoint in readily erodible material, the knickpoint would be obliterated rapidly. Knickpoints exist in the field at all stages between these extremes. Although partly a function of factors pre-

viously mentioned, they are controlled primarily by the erodibility of the bed material.

In an ideal case in which a knickpoint has been formed on the bed of a channel of well-sorted noncohesive material, certain changes may be expected in the longitudinal profile provided that the flow conditions cause erosion of the channel bed. As erosion along the oversteepened reach is a function of the transport rate at any point, it would be desirable to express the relative rates of erosion along the knickpoint reach as some function of the flow parameters that govern the transport rate. Although many empirical and partially theoretical equations have been proposed to describe the rate of bed-load movement, agreement is not complete. Nevertheless, many workers agree that shear forces acting along the boundary are of primary importance in streams containing nearly plane beds:

$$q_s = f(\tau_0) \tag{1}$$

where q_s equals the rate of transport per unit width, τ_0 is the shear at the bed, and f is a function. The shear at the bed is commonly expressed as

$$\tau_0 = \gamma d S \tag{2}$$

where γ is the specific weight of water, d the depth, and S the slope of the energy profile. Therefore, q_s varies along the oversteepened reach if the depth-slope product does not remain constant.

Many combinations of bed slopes accompany knickpoints, but one of the more common combinations is that of a gentle reach upstream from a knickpoint and a rather steep downstream reach. In hydraulic terminology, the upstream reach is usually mild, and the downstream reach is steep. A mild slope is one that will sustain uniform flow only at subcritical velocities, and a steep slope will sustain uniform flow at supercritical velocities (Rouse, 1949, p. 610). The division of velocity into sub-

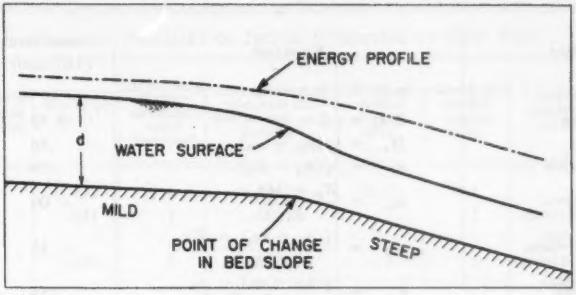

FIGURE 1.—FLOW OVER AN OVERSTEEPENED REACH
The erosion potential reaches a maximum at the break in slope and decreases away from this point.

critical or supercritical depends on the Froude number:

$$F = \frac{v}{\sqrt{gR}} \tag{3}$$

where v is the mean velocity, g is the acceleration due to gravity, and R is the hydraulic radius which is approximately equal to the depth for wide shallow channels. The velocity associated with a Froude number less than 1 is subcritical (streaming turbulent flow), and the velocity accompanying a Froude number greater than 1 is supercritical (shooting turbulent flow).

To study the changes that occur in the transition from mild to steep slopes, it is necessary to derive the backwater curves for the zone of study (Rouse, 1949, p. 611). With these curves, it is possible to observe the change in shear at the bed along the knick-point reach by plotting the change in the depth-slope product. Figure 1 shows the bed, water-surface, and energy profiles that occur in the transition from a mild to steep bed slope.

Just at the knickpoint, the depth-slope product reaches a maximum, indicating that the boundary shear has a maximum value. This, in turn, suggests an extremely large rate of sediment transport. Downstream from this point, the depth decreases and the slope increases, but the resulting boundary shear decreases downstream from the knickpoint. Upstream from the

knickpoint, the depth increases, and the slope of the energy profile decreases. The resulting dS product decreases upstream from the knickpoint because the energy profile flattens much more rapidly than the depth increases. Therefore, at some distance upstream from a knickpoint, the erosion potential represented by the shear is relatively small but increases downstream, reaches a maximum directly over the knickpoint, and then decreases downstream. Assuming that the amount of sediment transport is proportional to the shear on the bed, more material will be carried away from the site of the knickpoint than from the reaches upstream or downstream. Therefore, after a very short time, the position of the knickpoint will migrate upstream. Owing to the fact that the transport decreases in a downstream direction, the tendency would also be to decrease the slope of the oversteepened reach below the knickpoint. Finally, the slight erosion occurring upstream from the knickpoint would tend to reduce the abruptness of the slopes adjacent to the knickpoint as it migrates upstream. The process is repeated during succeeding time intervals, although the magnitude of erosion decreases as the reach below the knickpoint flattens.

In some places other combinations of bed slopes, such as mild to mild and steep to steep, bound a knickpoint. However, the erosion potential and the method of migration of the knickpoint in addition to the accompanying reduction in slope of the reach below the knickpoint proceed as was discussed in the previous analysis. This occurs despite the fact that different backwater curves must be used to describe the water-surface profiles for the various combinations of bed slopes.

Thus, over a period of time the slope of the oversteepened reach below a knickpoint in homogeneous bed material will decline if the prevailing flows are capable of moving the bed material. Factors such as changes in channel shape and irregular deposition of the eroded material might alter the rates of change of slope, but in the absence of resistant material, a reduction in slope should occur.

TABLE 1.—SIZE CHARACTERISTICS OF BED MATERIAL

Statistic	Equation	Coarse sand (0.67 mm)	Very coarse sand (2.0 mm)
Phi median	$Md_\phi = \phi_{50}$	0.58	−1.0
Phi mean	$M_\phi = \frac{1}{2}(\phi_{16} + \phi_{84})$	.56	−1.0
Phi deviation	$\sigma_\phi = \frac{1}{2}(\phi_{84} - \phi_{16})$	.26	.50
Phi skewness	$\alpha_\phi = \frac{M_\phi - Md_\phi}{\sigma_\phi}$	−.08	0
2d phi skewness	$\alpha_{2\phi} = \frac{\frac{1}{2}(\phi_5 + \phi_{95}) - Md_\phi}{\sigma_\phi}$	−.15	−.12
Phi kurtosis	$\beta_\phi = \frac{\frac{1}{2}(\phi_{95} - \phi_5) - \sigma_\phi}{\sigma_\phi}$	.54	.64

where $\phi = -\log_2$ (mm)

Equipment and Procedure

The flume operated by the U. S. Geological Survey at the University of Maryland, College Park, Maryland, is 52 feet long and 4 feet wide. It is supported on a double truss and can be tilted to slopes as high as 3 per cent. Sediment can be fed into the upper end of the flume from a hopper by means of a vibrating feeder. The outflow of sediment is measured in a weighing tank at the lower end of the flume. Water discharge is circulated through a sump and constant head tank and is measured by an elbow meter located above a stilling basin at the entrance to the flume.

Before each run, a trapezoidal channel 0.7 foot wide and about 0.1 foot deep with rounded corners was molded in the noncohesive sand. The first three runs were made with a sand 0.67 mm in diameter and the last two with a sand 2.0 mm in diameter. The population characteristics of the two sands are given in Table 1. For each of the runs, a reach 1 foot long between stations 35.5 to 36.5 was oversteepened to provide a fall of approximately

0.1 foot. Station 0 is at the head of the flume, and station 52 is at the exit. The slopes of the channels above and below the oversteepened reach, here called the average slope, were approximately equal in each run, but the average slope for the upstream and downstream reaches was varied in different runs.

Initial Conditions of Test Runs

The starting conditions for the five runs in noncohesive material are shown in Table 2.

TABLE 2.—SUMMARY OF INITIAL CONDITIONS OF TEST RUNS

Run no.	Median size of bed and bank material (mm)	Discharge (cfs)	Initial slope of oversteepened reach (feet/feet)	Average slope of upstream and downstream reach (feet/feet)	Width of molded channel (feet)	Depth of molded channel (feet)	Sediment transport in reaches above and below oversteepened reach
1	0.67	0.021	0.1	0.00125	0.7	0.1	none
2	.67	.021	.1	.00315	.7	.1	moving estimated at .000035 lbs/sec.
3	.67	.030	.1	.00310	.7	.1	moving estimated at 00040 lbs/sec.
4	2.0	.021	.1	.00120	.7	.1	None
5	2.0	.032	.1	.00880	.7	.1	None

The longitudinal profile of the dry bed of the molded channel was measured with a point gauge mounted on a moving carriage. The water was then turned on at a constant discharge. The locations of the knickpoint and of the front of the dune that invariably formed below the oversteepened reach were recorded at successive times The water-surface slope was measured from time to time during each run. Early in the runs, the changes at the oversteepened reach were so rapid that only the water-surface slope in the immediate reach could be measured. As the process slowed down, water-surface slopes were measured over the entire length of the flume. Changes in width were also measured at various stations as the run progressed. At the completion of a run, a longitudinal profile of the dry bed was again recorded, and channel cross sections were measured. The duration of the runs ranged from 150 to 1500 minutes.

The selection of sand sizes and initial average slopes for the various runs was not arbitrary but was based on previous experimental results obtained by the authors for equilibrium channels in noncohesive materials (Wolman and Brush, in press). For example, runs 2 and 3 were made in equilibrium channels which would have retained their initial shapes for the given flow and sediment transporting conditions even without the presence of a knickpoint. Runs 1, 4, and 5 were made in stable channels in which the flow intensity was made insufficient for sediment motion except in the oversteepened reach.

As each run progressed changes took place in the bed and water-surface slope, the position of the knickpoint or break in slope, the position of the dune front below the knickpoint, and the shape of the channel. A record of these changes was made.

RESULTS

Changes in Slope

Both the slope of the water surface and the slope of the bed in the oversteepened reach decreased with time. Although changes along the bed cause changes in the slope of the water surface, the bed is much more difficult to measure without disturbing the flow. The water was turned off once during run 1, so that the bed profile might be measured at an intermediate stage as well as at the beginning and end of the run. Changes in the profile of the bed during run 1 are shown in Figure 2. At the beginning of this run no sediment was moving in the reaches above and below the oversteepened reach. The erosion and deposition noted in the figure, therefore, resulted solely from the influence of the knickpoint. Erosion was most vigorous during the first 2 hours and 40 minutes. Erosion above the knickpoint and deposition below contributed to the general decrease in gradient of the bed as the head and the toe of the original oversteepened reach moved upstream and downstream respectively. The

steep upstream slope at the end of the run is caused by the flume entrance which is fixed and cannot be eroded. The bed was much more irregular below the toe because of the formation and migration of the dune front. The profile

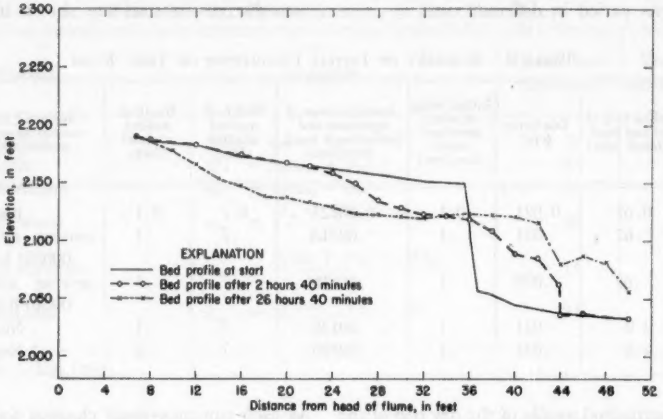

FIGURE 2.—CHANGES OCCURRING IN THE LONGITUDINAL PROFILE OF THE BED WITH TIME (RUN 1)

shows only two dimensions of the channel, though changes in width did occur, as discussed below.

The initial and final longitudinal profiles of the bed in each of the other runs are shown in Figure 3. In general, the changes in each run were similar, but with coarse material in runs 4 and 5 the rate of sediment transport was so slow that the bed irregularities below the knickpoint were very prominent.

Water-Surface Slope

The slope of the water surface in the over-steepened reach below the original knickpoint changed concurrently with changes in the bed profile. The ease of measurement of the water surface enabled a fairly detailed record to be made throughout each run. In all runs the slope of the water surface in the vicinity of the over-steepened reach decreased with time. Successive water-surface profiles for run 1 are shown in Figure 4. These profiles are smoother and more regular than the bed profiles (Fig. 2). The profiles appear to pivot around a horizontal axis lying 2–3 feet from the original over-steepened reach. This same pivotal change in water-surface slope was found also in the other runs.

To visualize the change in water-surface slope for each of the runs a summary graph in which water-surface slope is plotted against time for all five runs is given in Figure 5. This plot shows that for the 0.67-mm sand there is no difference in the change of slope with time between the runs despite the fact that in run 3 the discharge was increased by about 50 per cent, and in runs 2 and 3 the average slope of the reaches above and below the knickpoint was nearly trebled. Slope and discharge were not varied sufficiently, however, to cause a major change in the fitted curve of slope against time. For the runs in the 2.0-mm sand, the average slope does influence the position of the plotted points on the curve. The large average initial slope in run 5 caused a greater total reduction in slope near the knickpoint than occurred in run 4. The average initial slope for run 4 was 0.00120, whereas the slope was 0.00880 for run 5. These results show that the rate of change of the slope at the knickpoint is influenced by differences in average slope but is not particularly sensitive to them.

Particle size seems to have a very strong influence on the rate of change of slope with time. For example, runs 1 and 4 are nearly identical except for particle size. However, the per cent change in slope in the finer material is much greater for any given unit of time than it is in the coarser material.

Change in Slope with Time at Different Reaches

The relations discussed previously involved the slope over the entire length of the over-

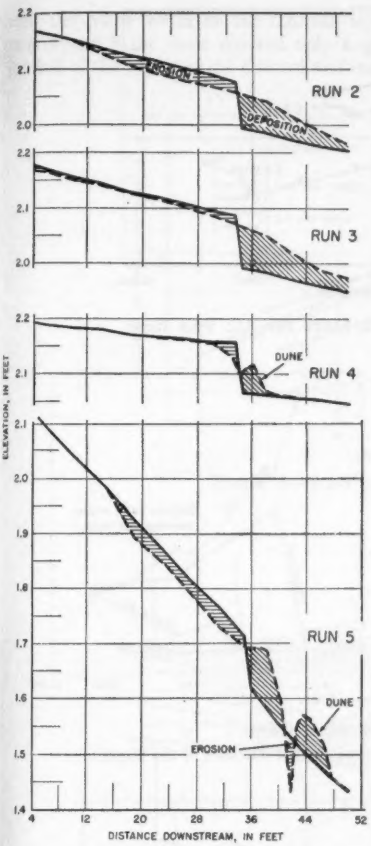

FIGURE 3.—STARTING AND FINAL BED PROFILES OF RUNS 2, 3, 4, AND 5

The source of sediment for deposition was not the upstream bed alone but also the channel sides as a result of widening. Sand was fed to the channel during runs 2 and 3.

steepened reach. Successive changes in slope that take place at each cross section are also important in describing the characteristics of knickpoint migration. These are shown for run 1 in Figure 6 for sections above and below the original position of the knickpoint. Upstream from the knickpoint the slope gradually increases at each section until a maximum is reached, after which the slope declines. The result is that after a long interval, the channel will become adjusted to a slope that is nearly equal throughout its length and not much different from the original slope. For example, starting with a slope of about 0.00125 in run 1, the passage of the knickpoint and the readjustment of the channel in about 1600 minutes returned the slope to about 0.0030, and the general trend of the curves suggests that the slope would continue to decrease. Because of the limited length of flume, the process of slope reduction could not go to completion.

Downstream the general changes in water-surface slope at each section tend to be similar to the upstream changes in that slope first increases to a maximum and then decreases. In the lower reach, however, steepening occurs by deposition on the bed rather than by erosion.

Growth of the Oversteepened Reach

Measurements were also made of the rate of retreat of the knickpoint and of the rate of advance of the dune front (Figs. 7, 8). As the longitudinal growth of the oversteepened reach determines the change in the water-surface slope, and the growth of the dune front depends in part on the rate at which sediment is contributed by erosion above the knickpoint, an agreement between the rates of change of each of these variables should exist.

Comparison of Figures 7 and 8 with Figure 5 does show that, with the exception of run 5, the

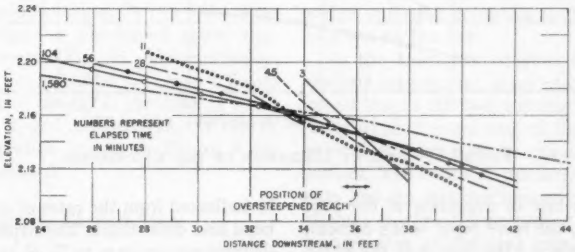

FIGURE 4.—CHANGE IN WATER-SURFACE SLOPE FOR RUN 1

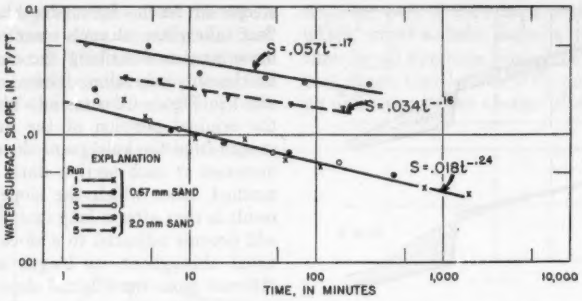

FIGURE 5.—CHANGE IN WATER-SURFACE SLOPE FOR ALL FIVE RUNS

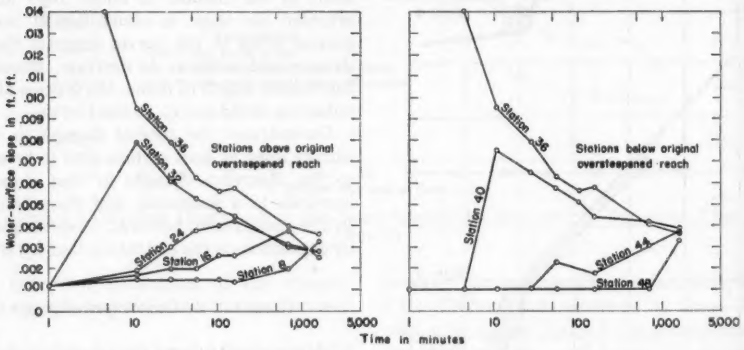

FIGURE 6.—CHANGES IN SLOPE FOR RUN 1
Station numbers represent the distance in feet from the entrance of the flume.

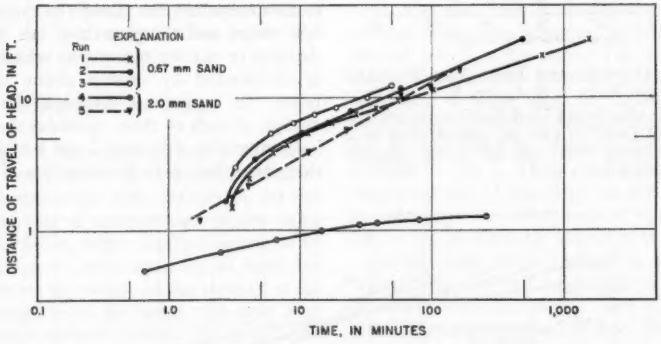

FIGURE 7.—RATE OF MIGRATION OF THE KNICKPOINTS

more rapid the rate of migration of the head and dune front, the more rapid is the decrease in slope. During run 5 the change in the water-surface slope occurred more slowly than would be indicated from the rates of movement of the head and dune front. The explanation for this discrepancy appears to lie in the nature of the dune front. In runs 1, 2, and 3, the dune cov-

ered the entire width of the channel, but in runs 4 and 5 the dune covered only a small portion of the center of the channel and caused caused by erosion and deposition along the channel. In channels that have noncohesive banks composed of well-sorted sand, like those

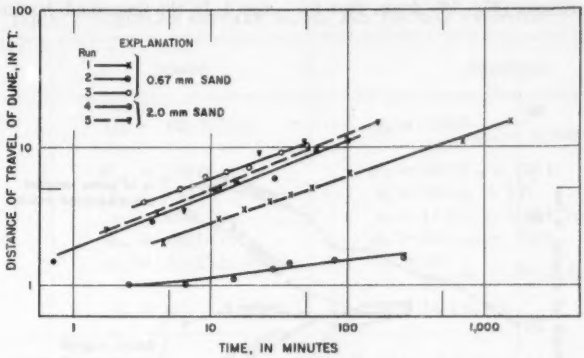

FIGURE 8.—RATE OF MIGRATION OF DUNE FRONTS

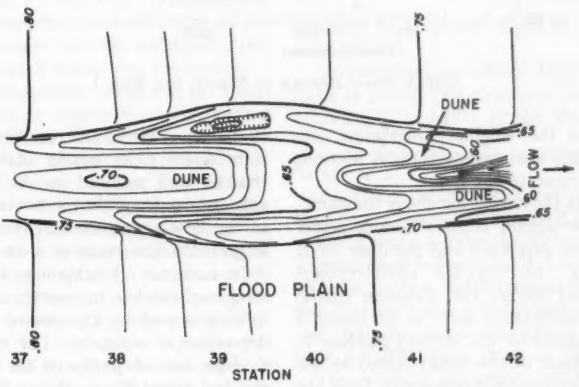

FIGURE 9.—CONTOUR MAP OF CHANNEL AT THE DUNE FRONT, RUN 5
Channel formed in 2.0-mm sand. Contour interval is 0.01 feet.

the flow to divide, producing a pattern similar to that of a braided stream. In run 5 the dune front is a wedge- or pie-shaped prism that tapers downstream (Fig. 9). Volumetrically, less sediment was required to cause a rapid rate of advance of the tip of the dune in run 5 than in runs 1, 2, and 3. Therefore, the rate of dune-front migration shown in Figure 8 for run 5 is exaggerated.

Changes in Channel Shape

The previous analyses of run 5 indicate that shape changes accompany changes in profile used in this study, the channel was free to widen or deepen according to the requirements imposed by the flow.

As the knickpoint migrated upstream, the channel was eroded most strongly along the center line of the bed, causing a relative deepening of the central part of the channel and a general narrowing of the channel at the water surface. The width of the band of central erosion and its relative intensity decreased upstream, so that the deepening and narrowing became less pronounced with distance away from the original position of the knickpoint. The width of the band of erosion and the degree

of entrenchment decreased upstream because the over-all slope of the reach after the passage of the knickpoint decreased with time of run. A plot of the changes in width with time at various cross sections (Fig. 10) shows that for dune formation, and division and reconcentration of flow extended downstream with decreasing intensity. A similar though less well-developed channel configuration occurred during run 4. In the finer sand, however, the distribu-

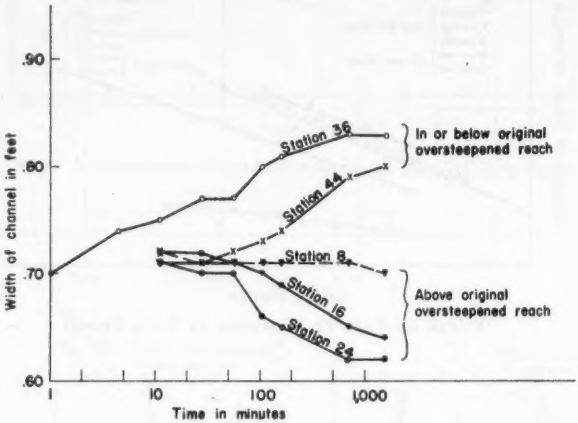

FIGURE 10.—CHANGES IN WIDTH FOR RUN 1

sections above the original knickpoint, decreases in width are smaller for the stations nearer the entrance to the flume.

In contrast to the narrowing above the knickpoint, the downstream reaches became wider as sediment was deposited and the dune front advanced (Fig. 10). In the oversteepened reach itself and below, the widening results from the increase in shear forces at the banks of the channel caused by the marked increase in slope. The erosion of the banks added to the sediment being brought downstream from the erosion of the head and tended to increase the rate of advance of the dune front.

The configuration of the channel also depends in part on the ability of the stream to transport the size of material present in the bed. In run 5 with 2-mm sand, sediment was moved into the dune only in the center of the channel, seemingly because the shear along the edges of the channel was not sufficient to transport the sediment required to form the dune across the entire width of the channel. As a result, the central dune caused the flow to divide and scour the bed on both sides. Downstream from these lateral points of scour, small secondary dunes were formed at the sides of the channel. These in turn reconcentrated the flow in the center of the channel. This succession of scour, tion of shear forces and resulting transportation were much more nearly uniform across the channel and resulted in the formation of a single dune front that extended perpendicular to the flow across the full channel width and migrated downstream as a unit.

In summary, knickpoints and short oversteepened reaches in noncohesive channels are quickly erased by the process of erosion and deposition of sediment. The rate of reduction in slope depends partly on the size of the material but primarily on the magnitude and rate of transport along the oversteepened reach which, in turn, governs the rate of retreat of the knickpoint at the head and concomitant building of the dune below. As the longitudinal profile changed, channel shape also changed, depending upon whether the reach was undergoing degradation or aggradation. Above the original knickpoint, the channel became deeper and narrower, whereas below the knickpoint it became shallower and wider.

DISCUSSION

Comparison with Other Model Studies

Others have made flume studies of knickpoints, but with the exception of Lewis's (1944)

almost all experimental studies have been concerned with natural meander cutoffs (U. S. Waterways Experiment Station, 1932). For the most part, few data are available from the model cutoff studies on progressive changes in water-surface slope of the cut-off or on the detailed changes in the bed. Qualitatively, water-surface slope decreased with time as the bed eroded upstream from the knickpoint and as the bed aggraded below the knickpoint.

Lewis shows progressive changes in the longitudinal profile in an experimental run lasting 17 hours and 49 minutes. The data show that the slope below the knickpoint increased during the first 19 minutes, but that after 49 minutes it decreased to a value less than the original. As the run continued, the slope increased slightly after 9 hours and 49 minutes but decreased again after 17 hours and 49 minutes. The interpretation of this experiment is complicated, however, by the presence of two tributaries in the same experimental channel and by the fact that the bed material was composed of a mixture of "4 parts sand to 1 part mud." The changes in slope below the knickpoint were complicated by flow from the tributaries, but an over-all decrease in slope occurred until practically all traces of the original oversteepening were gone. The initial increase in slope is of interest because this differs from the results obtained in the authors' experiments. Possibly, the original knickpoint was near a pocket of more resistant "mud," which temporarily inhibited erosion upstream from the knickpoint. Lewis's (1944) results agree otherwise with the authors'.

Prototype Equivalent to the Model Study

To give some idea of a comparison between small model streams and large natural streams an attempt is made below to estimate the characteristics of the larger stream using the principles of dimensional similitude. Run 1 is used in the analysis as a typical model run.

TABLE 3.—RELATION BETWEEN MODEL AND POSSIBLE PROTOTYPE

Parameter	Model	Prototype
Discharge	$Q_m = .021$ ft.³/sec.	$Q_p = \frac{(200)^2}{14.1} Q_m = 11{,}900$ ft.³/sec.
Width	$W_m = .70$ ft.	$W_p = 200\, W_m = 140$ ft.
Depth	$d_m = .045$ ft.	$d_p = 200\, d_m = 9$ ft.
Velocity	$v_m = .67$ ft./sec.	$v_p = 14.1\, v_m = 9.45$ ft./sec.
Particle size	$d_{sm} = .00219$ ft.	$d_{sp} = 200\, d_{sm} = .44$ ft.
Bed slope	$S_m = .00125$ ft./ft.	$S_p = \frac{200}{200} S_m = .00125$ ft.
Time	$t_m =$ sec.	$t_p = 14.1\, t_m$ sec.

(1) Because the Reynolds number of the flow, $\frac{4Rv}{\nu}$, was moderately high (11,200) it was assumed that viscous effects in the model are unimportant.

(2) Gravitational effects, therefore, form the basis of model and prototype similarity. This criterion is satisfied if the Froude number, $\frac{v}{\sqrt{gR}}$, is the same in model and prototype.

(3) The geometric scale ratio of the prototype to the model was assumed to be 200 to 1.

The Froude number in the model was 0.59. The relations between model and prototype are given in Table 3, which is based on the assumption that the value in the prototype is the same and that the length dimensions in the prototype are 200 times those in the model.

The resulting prototype stream is somewhat similar to streams such as those on outwash plains in Alaska with steep slopes, coarse bed material, and high mean velocities. Some distortion appears, however, as the width is somewhat narrower and velocity somewhat higher than would be expected for average streams with a discharge of 11,900 cfs (Leopold and Maddock, 1953, p. 15). Despite these qualifications, it is interesting to extend the estimation of the prototype characteristics further in light of the flume results.

A knickpoint will stop moving for any of three reasons: (1) if the reduction in slope of the oversteepened reach reduces the competence of the stream to the point where it cannot move the material on the bed, the knickpoint will cease migrating upstream; (2)

the knickpoint may reach a control such as a dam or a very resistant outcrop in the bed and become stabilized at that point; (3) a knickpoint may become indistinguishable from its surrounding bed by lengthening of the oversteepened reach to the point where its slope is nearly equal to the slope of the stream. A variation on this last example might be that after the cutoff has formed and the oversteepened slope has begun to flatten, the oversteepened reach may begin to lengthen its course by meandering, which would cause further reduction in the slope of the oversteepend reach. All these alternatives may occur in natural streams although the first, reduction in competence, is the least likely. In a flume study, competence and flume length may be the limiting factors.

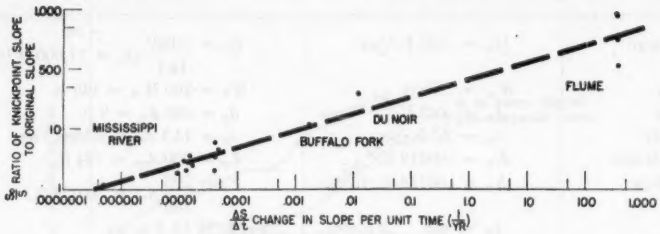

FIGURE 11.—RELATION BETWEEN RATIO OF SLOPE OF OVERSTEEPENED REACH TO AVERAGE SLOPE AGAINST RATE OF CHANGE OF SLOPE

Because competence and natural or artificial controls may be important in determining the rate of migration of a knickpoint as well as of the rate of flattening of the oversteepened slope, the maximum distance over which a knickpoint may travel, given unlimited time, is essentially a problem in geometry. The greater the ratio of the slope of the oversteepened reach to the average slope of the channel, the farther a knickpoint will migrate upstream.

If no controls occur within the reach, the oversteepened reach will tend to flatten asymptotically toward the average slope of the channel. The writers assume for this discussion that a knickpoint cannot be identified if the over-steepened reach flattens to a slope within 20 per cent of the average slope of the river. For the flume experiments it may be shown by simple geometry that if the original average slope is 0.00125, if the oversteepened reach has a fall of 0.1 foot in 1.0 foot, and if the head and toe of the oversteepened reach move at approximately the same rates, then the oversteepened reach will flatten to 0.00150 (20 per cent of the average) after the knickpoint has migrated 33 feet. In a prototype stream that has the same geometry and a fall of 20 feet in 200 feet, the oversteepened reach would return to within 20 per cent of the average slope after the knickpoint had migrated 6600 feet.

Any scale ratio of prototype to model less than the 200 to 1 assumed in the foregoing analysis would cause the knickpoint to become indistinguishable after a shorter distance of travel. A larger scale ratio of 2000 to 1 would yield a migration distance of 66,000 feet for the knickpoint. However, this scale ratio would require also a width of 1400 feet (which is comparable to that of the Mississippi River at St. Louis, for example) and would require an unrealistically high initial oversteepening of 200 feet in 2000 feet. These examples of hypothetical prototype streams show that a knickpoint probably would not be recognizable more than several miles upstream from its initial position unless the original fall was extremely high; it would be most unlikely to travel the entire length of a natural stream.

As in most natural streams, the knickpoint or the dune front may encounter a control during the migration process. If this occurs, the oversteepened slope will not return to the average slope but will remain at some higher value. In the flume, the ends would be the controls. This resulting slope might be called the "trend slope" (Shulits, 1955, p. 651). Trend slopes might be reached in natural streams because of numerous types of controls as previously mentioned.

Cutoffs in Natural Rivers

Detailed information similar to that obtained in the flume is lacking for progressive changes at natural knickpoints. Figure 11 shows a plot of all the data available to the authors and in

addition to the flume data includes actual cutoffs in 11 reaches along 3 rivers. The wide range of slopes represented—from 0.000021 in the Mississippi River to the neighborhood of 0.001 in the flume channels—is shown in Table 4.

TABLE 4.—PROGRESSIVE CHANGES IN SLOPE AT MEANDER CUTOFFS IN NATURAL RIVERS

Stream	Ratio of slope of oversteepened reach to average slope	Change in slope per year	Source of data
Mississippi River at Ashbrook cutoff	4.7	0.000047	U. S. Mississippi River Commission (1940)
Worthington	1.2	.0000005	U. S. Mississippi River Commission (1940)
Sarah	3.5	.000041	U. S. Mississippi River Commission (1940)
Willow	2.5	.000017	U. S. Mississippi River Commission (1940)
Marshall	2.1	.000013	U. S. Mississippi River Commission (1940)
Diamond	6.1	.000041	U. S. Mississippi River Commission (1940)
Yucatan	3.6	.000015	U. S. Mississippi River Commission (1940)
Rodney	1.9	.000009	U. S. Mississippi River Commission (1940)
Giles	4.1	.000021	U. S. Mississippi River Commission (1940)
Glasscock	3.0	.000012	U. S. Mississippi River Commission (1940)
DuNoir Creek (Wyoming)	39.5	.0121	Field observations
Buffalo Fork (Wyoming)	12.1	.00177	Field observations
Flume run 1	800	368	This study (laboratory)
Flume run 2	318	368	This study (laboratory)
Flume run 3	323	360	This study (laboratory)
Flume run 4	833	361	This study (laboratory)
Flume run 5	114	364	This study (laboratory)

The data of Macar (1934), whose work on cutoffs was excellent, could not be included in Figure 11, because he did not collect quantitative measurements of the rate of change of slope. Lane (1947) describes the effects of cutoffs in erodible and non-erodible channels. His general observations confirm the experimental results found in the model study, but no additional quantitative data are provided.

The ratio of the oversteepened slope below the knickpoint to the average slope is an index of the relative magnitude of the disturbance caused by the cutoff. Hence, a high ratio of S_0/S indicates that a high relative rate of transport of bed material will occur and, according to the previous discussion, should be coincident with a rapid change in slope at the cutoff. Figure 11 seems to verify this reasoning.

Geologic Implications

In the flume study the slope below the knickpoint decreased with time, owing to deposition. The rate of reduction of slope was shown to be a function of the rate of transport along the oversteepened reach. Under the same flow conditions, the change of slope with the 2.0-mm sand was much less than it was with the 0.67-mm sand. In nature, however, vertical cliffs such as small headcuts and falls such as Niagara Falls may migrate upstream without a decrease in slope. These are special cases,

and the discussion below attempts to show that they are consonant with the original hypothesis that the rate of change of slope is a function of the transport and erosion at the knickpoint. However, no attempt is made here to classify or generalize the vast geologic literature on knickpoint formation and behavior.

FIGURE 12.—SCHEMATIC DRAWING OF THE INFLUENCE OF VARIOUS COMBINATIONS OF RESISTANT AND NONRESISTANT BED MATERIAL ON KNICKPOINT CHANGES

Numerous other authors have described many or all of the conditions considered by the present authors. The relatively simple experimental conditions, however, seemed to provide a useful basis for discussing the invariably more complex cases in nature.

For the purposes of discussion, five hypothetical examples of different geologic environments are considered (Fig. 12). Type A is similar to the flume experiment and on the basis of these experiments should apply to all knickpoints in homogeneous materials, *i.e.*, large alluvial valleys. In their case, as shown before, the final slope will approach the average channel slopes. Type B has a resistant stratum intersecting the profile at an arbitrary angle and represents a typical case which might be encountered in any region of folded sedimentary rocks. Here a knickpoint will retreat and reduce its slope until the resistant layer is intersected. At this time, the rate of erosion of the stream is decreased in the nonresistant material, which will tend to preserve a steeper slope. This break will continue indefinitely or until the resistant layer is removed, because it always represents an area of slow erosion. If the potential for erosion downstream is adequate, a falls (type C) might develop to the extent that material can be removed on the decreasing slope below. If all traces of the resistant strata are removed condition B becomes identical with A. In nature, type B is perhaps the most general case.

Type C, the Niagara Falls case, occurs where a resistant bed caps an underlying nonresistant bed. In this example the slope below the knickpoint may increase toward the vertical depending on the relative differences in resistance to erosion and in transport of the two kinds of bed material.

Types D and E are essentially possible variants of the three preceding examples. Knickpoint type D would behave similarly to one in homogeneous material if the resistant layer lay below the center of the oversteepened reach and dipped downstream at an angle less than that of the final or average slope. Downstream case B will occur if the resistant strata are uncovered. Finally, type E, a variation of D, shows the influence of a resistant layer buried only slightly below the surface but exposed at the oversteepened reach. It is essentially an extension of D.

Several conclusions are indicated: (1) If a stream has homogeneous bed material and is capable of transporting this material easily, knickpoints would disappear quickly. (2) Resistant material near the original site of a knickpoint would tend to preserve the knickpoint and localize the change in slope. However, inasmuch as the slope of the stream depends on the relative ability of the stream to transport and erode its bed in successive reaches, a break in the longitudinal profile would be expected at the point of change in bed material with or without the prior initiation of a knickpoint downstream. (3) A knickpoint will not easily bypass a resistant layer. Hence, a large number of knickpoints caused by uplift, changes in base level, or climatic changes may all migrate and merge at one point along a stream profile. (4) Falls are a special case, as described by many authors, in which the steep slope is caused by undermining. Headcuts in alluvium or colluvium may or may not fit into

this category. Seepage or underflow as well as differential resistance may be important in headcut preservation and migration.

Conclusion

The model study of knickpoint migration and associated slope changes in noncohesive, well-sorted sands permits some generalizations about natural or field conditions. Although backed up in the laboratory by quantitative data, they can be extended only qualitatively to the field. Bed-load movement is a prerequisite for producing changes in a knickpoint once it has formed. Sediment-transport principles suggest that an oversteepened slope in homogeneous material bounded by less steep slopes upstream and downstream tends to lessen in gradient with time. Changes in channel shape accompany profile changes. Upstream the channel tends to become narrower and deeper, whereas downstream from the original oversteepened reach the stream tends to become wider and more shallow. Because of the tendency for a stream to reduce rapidly any oversteepened reach in homogeneous bed material, the knickpoint migrates upstream only a short distance before becoming too faint to recognize. At a given flow the rate of migration will depend on the size of the material, and the total distance of migration will depend on the difference in fall between the average and oversteepened reach. Introducing more resistant material into the stream may cause local oversteepening under some circumstances. Unless the total fall increases the knickpoint will not migrate upstream any farther than it would in homogeneous material but will do so at a much slower rate. Unless the resistant material is removed the profile will be steeper because of the resistant material, which will tend to be the temporary locus of any additional knickpoints that might form immediately downstream.

References Cited

American Geological Institute, 1957, Glossary of geology and related sciences: Natl. Acad. Sci.-Natl. Research Council, Pub. 501, 325 p.

Lane, E. W., 1947, The effect of cutting off bends in rivers: 3rd Hydraulics Conf. Proc., Bull. 31, Univ. Iowa, Studies in Engineering, p. 230–240

Leopold, L. B., and Maddock, T., 1953, The hydraulic geometry of stream channels and some physiographic implications: U. S. Geol. Survey Prof. Paper 252, 55 p.

Lewis, W. V., 1944, Stream trough experiments and terrace formation: Geol. Mag., v. 81, p. 241–253

Macar, Paul F., 1934, Effects of cutoff meanders on the longitudinal profiles of rivers: Jour. Geology, v. 42, p. 523–536

Rouse, Hunter, 1949, Engineering hydraulics: N. Y., John Wiley & Sons, Inc., 1039 p.

Shulits, S., 1955, Graphical analysis of trend profile of a shortened section of river: Am. Geophys. Union Trans., v. 36, p. 649–654

U. S. Mississippi River Commission, 1940, History of the improvement of the lower Mississippi River for flood control and navigation, 1932–1939: Corps of Engineers, U. S. Army, 198 p.

U. S. Waterways Experiment Station, 1932, Experiment to determine the effects of proposed dredged cutoffs in the Mississippi River: Paper I, 62 p.

Von Engeln, O. D., 1949, Geomorphology: N. Y., The MacMillan Co., 655 p.

Wolman, M. G., and Brush, L. M., Jr., in press, Experimental study of factors controlling the size and shape of stream channels in coarse noncohesive sands: U. S. Geological Survey Prof. Paper

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CLAYS AND CLAY MINERALS OF NEW ENGLAND AND EASTERN CANADA

BY VICTOR T. ALLEN AND W. D. JOHNS

ABSTRACT

Clay minerals were collected from shales and clay deposits in New England and Eastern Canada ranging in age from Precambrian to Recent. The collection was made to include representative clay minerals of each district, geological period, and environment. Fifty-six samples were selected for detailed petrographic and X-ray study. The potash clay minerals, or hydrous micas, and chlorite are the abundant clay minerals in the marine shales examined. The hydrous micas, chlorite, or vermiculite are the abundant clay minerals in the varved clays of nonmarine origin and clays containing marine fossils of Quaternary age. Clay from glacial till also contains potash clay minerals and vermiculite. The source material had a greater influence on the clay-mineral composition than the environment in which the clay minerals were deposited. The clays and shales that contain abundant kaolinite were apparently deposited under nonmarine conditions. They were eroded from kaolinite clays formed by surface weathering or by hydrothermal replacement. Kaolinite deposits attributed to these processes occur in this region.

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INTRODUCTION

This investigation of the clay minerals in the clays and shales of New England and Eastern Canada was undertaken with a threefold purpose: (1) to extend to this region (Fig. 1) the clay-mineral studies carried on by the writers over a period of years at many localities in the United States; (2) to examine the evidence that certain clay minerals can be used as indicators of special environmental conditions in which they were deposited; (3) to determine whether clay minerals are helpful in deciding whether the sediment in which they are present originated under marine or nonmarine conditions.

The collection of samples was planned not only to obtain representative clay minerals of each district but also to obtain a distribution that would be representative of the clay min-

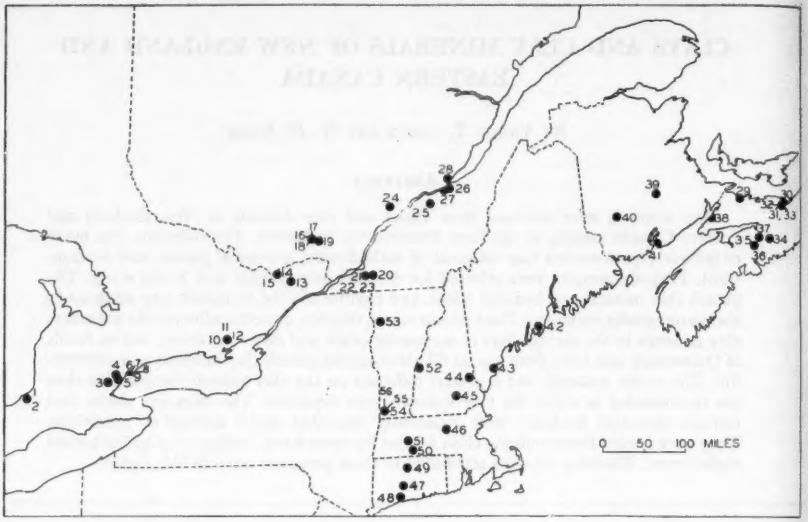

FIGURE 1.—INDEX MAP OF NEW ENGLAND AND EASTERN CANADA SHOWING LOCALITIES WHERE SAMPLES WERE STUDIED

1, 2. Hamilton shale, Devonian, W. S. Watson Pit, Thedford, Lambton County, Ontario
3. Queenston red shale, Ordovician, Milton Brick, Ltd., Milton, Ontario
4. Queenston red shale, Ordovician, F. B. McFarren, Ltd., Streetsville, Ontario
5. Dundas shale, Ordovician, Cooksville Co., Ltd., Cooksville, Ontario
6. Dundas shale, Ordovician, Don Valley Brick Works, Toronto, Ontario
7, 8. Clay, Quaternary, Don Valley Brick Works, Toronto, Ontario
9. Silty Clay, Quaternary, Toronto Brick Co., Greenwood Ave., Toronto, Ontario
10, 11, 12. Clay, Quaternary, Napanee Brick and Tile Co., Napanee, Ontario
13. Medina red shale, Silurian, Ottawa Brick and Terra Cotta Co., Russell, Ontario
14. Clay, Quaternary, Ottawa Brick and Terra Cotta Co., Ottawa, Ontario
15. Chazy Shale, Ordovician, Rockcliffe Park, Ottawa, Ontario
16, 17, 18. Kaolin, St. Remi D'Amherst, 7 miles from Huberdeau, Labelle County, Quebec
19. Clay, nonbedded, Quaternary, Huberdeau, Labelle County, Quebec
20. Clay, bedded, Quaternary, Standard Clay Products, Co., St. Jean, St. John County, Quebec
21. Boulder clay, Quaternary, National Brick Co., Delson Junction, Quebec
22. Utica shale, Ordovician, National Brick Co., Delson Junction, Quebec
23. Lorraine shale, Ordovician, Laprairie Brick Co., Laprairie, Quebec
24. Clay, Quaternary, Riviere Cachee, 14 miles north of Three Rivers, Quebec
25. Clay, stratified, Quaternary, Deschaillons, Quebec
26. Levis shale, Ordovician, Levis, Quebec
27. Charny shale, Lower Cambrian, Quebec City, Quebec
28. Utica-Lorraine shale, Ordovician, Montmorency Falls, Quebec
29. Red shale, Mississippian, Pugwash, Nova Scotia
30. Red shale, Pennsylvanian, Woodburn, Nova Scotia
31. Gray shale, Pennsylvanian, New Glasgow, Nova Scotia
32. Oil shale, Pennsylvanian, New Glasgow, Nova Scotia
33. Gray shale, Pennsylvanian, New Glasgow, Nova Scotia
34. Mottled clay, Upper Jurassic (?), Middle Musquodoboit, Nova Scotia
35. Red shale, Mississippian, Milford, Nova Scotia
36. Clay, Quaternary, Elmsdale, Nova Scotia
37. Clay, Upper Jurassic (?), Shubenacadie, Nova Scotia
38. Clay, Quaternary, River Hebert, Nova Scotia
39. Shale, Pennsylvanian, Chipman, New Brunswick
40. Clay, Quaternary, Frederickton, New Brunswick
41. Red clay, Quaternary, St. John, New Brunswick
42. Clay, Quaternary, Thomaston, Maine
43. Clay, Quaternary, Saco, Maine
44. Clay, Quaternary, Gonic, New Hampshire
45. Clay, Quaternary, Hooksett, New Hampshire

(*Continued on next page*)

erals of each geological period and each environment of sedimentation in the district. Allen examined by petrographic methods all samples collected. The 56 samples listed in Table 1 were examined by X-ray diffraction by Johns. X-ray-diffraction patterns of oriented aggregates of the less than 2-micron clay fraction and crushed unfractionated samples were made and interpreted by Johns. Both writers share the responsibility for the statements and conclusions in this paper.

Acknowledgments

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Minerals in Shales

Clay Minerals

The term shale is used here for a rock that has well-developed lamination and contains abundant clay minerals. The abundant clay minerals in shales are the potash clay minerals, or hydrous micas, and chlorite (Table 1).

Careful study of the diagnostic 10A diffraction maximum of these potash clay minerals, or hydrous micas, indicates that two types may be present in these shales and clays. In some patterns the maximum is sharp and symmetrical, and these micas are referred to here as Type I hydrous micas. In other patterns the maximum is diffuse and smeared out asymmetrically toward low angles. This feature indicates significant hydration. These micas are referred to here as Type II hydrous micas (Table 1).

A specimen of Charny shale (Table 1, 27) near the bridge south of Quebec City is unusual because it is composed of essentially pure hydrous mica Type I, whereas most shales contain appreciable amounts of quartz and other minerals. Osborne (1954) reports that other samples from this formation contain a variety of minerals. Chlorite is absent, and vermiculite and a mixed-layer mica-chlorite mineral are present in the red Queenston shale at Streetsville, Ontario (4). In the Levis shale at Levis, Quebec (26), vermiculite also appears in place of chlorite. Three shales of Pennsylvanian age, two from New Glasgow, Nova Scotia (31, 32), and one from Woodburn, Nova Scotia (30), contain some kaolinite in addition to chlorite and abundant hydrous mica.

Other Minerals

Quartz is an abundant mineral in almost every shale studied from this region. In some specimens the quartz grains are sufficiently large to be seen in thin sections; in others evidence of their presence is indicated in the X-ray pattern. Feldspars are present in minor quantities or are absent. Many of the shales are calcareous, and X-ray patterns indicate a dominance of calcite, but minor to moderate amounts of dolomite are present in some shales. X-ray patterns indicate that hematite is the coloring mineral in the red shales.

Conditions of Deposition

The shales sampled during this investigation include strata representing both marine and

46. Clay, Quaternary, Still River, Massachusetts
47. Red clay, Quaternary, Middletown, Connecticut
48. Red clay, Quaternary, North Haven, Connecticut
49. Clay, Quaternary, South Windsor, Connecticut
50. Clay, Quaternary, Chicopee Falls, Massachusetts
51. Clay, Quaternary, Northampton, Massachusetts
52. Clay, Quaternary, Lebanon, New Hampshire
53. Clay, Quaternary, Essex Junction, Vermont
54. Clay, Quaternary, Bennington, Vermont
55, 56. Clay, Tertiary, Bennington, Vermont; depth (55) 66 feet; (56) 155 feet

TABLE 1.—Minerals in Clays and Shales from Localities Shown in Figure 1

M—marine; N—nonmarine; H—hydrothermal; R—residual; G—glacial; A—abundant; B—moderate; C—minor; D—trace; MC—mica-chlorite

ROCK TYPE	DISTRICT	GEOLOGICAL PERIOD	TYPE OF DEPOSIT	HYDROUS MICA		CHLORITE	VERMICULITE	KAOLINITE	MONTMORILLONITE	MIXED LAYER	QUARTZ	FELDSPAR	CALCITE	OTHER
				I	II									
1 SHALE	THEDFORD, ONT.	DEVON.	M	A		A					A	D	B	
2 SHALE	THEDFORD, ONT.	DEVON.	M		A	A					A	D	C	
3 SHALE	MILTON, ONT.	ORDO.	M		A	B				MC	A		B	HEMATITE
4 SHALE	STREETSVILLE, ONT.	ORDO.	M		A		B			MC	A	C	C	HEMATITE
5 SHALE	COOKSVILLE, ONT.	ORDO.	M	A		A					A	D	C	
6 SHALE	TORONTO, ONT.	ORDO.	M		A	B	C				A	B	C	
7 CLAY	TORONTO, ONT.	QUAT.	N	A		A	C				A	B	B	DOLOMITE
8 CLAY	TORONTO, ONT.	QUAT.	N	A		A	C				A	C	C	DOLOMITE
9 CLAY	TORONTO, ONT.	QUAT.	N	A		A	C				A	B	C	DOLOMITE
10 CLAY	NAPANEE, ONT.	QUAT.	N		A		A				A	C	C	DOLOMITE
11 CLAY	NAPANEE, ONT.	QUAT.	M		B		A				A	C	D	DOLOMITE
12 CLAY	NAPANEE, ONT.	QUAT.	M		B		A				A	C	D	DOLOMITE
13 SHALE	RUSSELL, ONT.	SIL.	M		A	A	C				A	D	D	HEMATITE
14 CLAY	OTTAWA, ONT.	QUAT.	M		A	B	C			MC	A	D		AMPHIBOLE
15 SHALE	OTTAWA, ONT.	ORDO.	M		A	B	C			MC	A			
16 KAOLIN	ST. REMI, QUE.	PRECAM.	H	D				A			A			ZIRCON
17 KAOLIN	ST. REMI, QUE.	PRECAM.	H	D				A			A			ZIRCON
18 KAOLIN	ST. REMI, QUE.	PRECAM.	H	D				A			A			ZIRCON
19 CLAY	HUBERDEAU, QUE.	QUAT.	N		A		A			MC	A	C		AMPHIBOLE
20 CLAY	ST. JEAN, QUE.	QUAT.	M		B		A				A	D		
21 CLAY	DELSON JUNCTION, QUE.	QUAT.	G		B		A				A	B		
22 SHALE	DELSON JUNCTION, QUE.	ORDO.	M		A	A					A	D	D	
23 SHALE	LAPRAIRIE, QUE.	ORDO.	M		A	A				MC	A	D	D	DOLOMITE
24 CLAY	RIVIERE CACHEE, QUE.	QUAT.	N		A		A		B		A	C	D	AMPHIBOLE
25 CLAY	DESCHAILLONS, QUE.	QUAT.	N		A						A	D	C	
26 SHALE	LEVIS, QUE.	ORDO.	M		A		A				A	D		
27 SHALE	QUEBEC CITY, QUE.	CAMB.	M	A										
28 SHALE	MONTMORENCY FALLS, QUE.	ORDO.	M		A	B					A	D		
29 SHALE	PUGWASH, N.S.	MISS.	M	A		C					A			
30 SHALE	WOODBURN, N.S.	PENN.	N		A	C		A			A			
31 SHALE	NEW GLASGOW, N.S.	PENN.	N		A	C		B			A			
32 SHALE	NEW GLASGOW, N.S.	PENN.	N		A	C		B			A			
33 SHALE	NEW GLASGOW, N.S.	PENN.	M		A	A					A			
34 CLAY	MUSQUODOBOIT, N.S.	U. JUR.?	N	D				A			A			
35 SHALE	MILFORD, N.S.	MISS.	M	A		B					A		D	HEMATITE
36 CLAY	ELMSDALE, N.S.	QUAT.	M	A		A					A	D	D	
37 CLAY	SHUBENACADIE, N.S.	U. JUR.?	N	C				A	D		A			
38 CLAY	RIVER HERBERT, N.S.	QUAT.	M	A			A				A			
39 SHALE	CHIPMAN, N.B.	PENN.	M	A		A					A			
40 CLAY	FREDERICKTON, N.B.	QUAT.	M	A		A					A			
41 CLAY	ST JOHN, N.B.	QUAT.	M	A		A					A			
42 CLAY	THOMASTON, ME.	QUAT.	N	A			B				A	D		
43 CLAY	SACO, ME.	QUAT.	M	A		C					A	D		
44 CLAY	GONIC, N.H.	QUAT.	M	A		C					A	D		
45 CLAY	HOOKSETT, N.H.	QUAT.	N	A		C					A	D		
46 CLAY	STILL RIVER, MASS.	QUAT.	N	A		B				MC	A	D		
47 CLAY	MIDDLETOWN, CONN.	QUAT.	N	A		B					A	D	D	
48 CLAY	NORTH HAVEN, CONN.	QUAT.	N	A		C					A	D	D	
49 CLAY	WINDSOR, CONN.	QUAT.	N	A		B					A	C		AMPHIBOLE
50 CLAY	CHICOPEE FALLS, MASS.	QUAT.	N	A		C					A			AMPHIBOLE
51 CLAY	NORTHAMPTON, MASS.	QUAT.	N	A		B					A			
52 CLAY	LEBANON, N.H.	QUAT.	N	A		A					A	C	C	AMPHIBOLE
53 CLAY	ESSEX JUNCTION, VT.	QUAT.	N	A		A					A	D		
54 CLAY	BENNINGTON, VT.	QUAT.	N	A		A					A	D		
55 KAOLIN	BENNINGTON, VT.	TERT.	R	B				A			A			
56 KAOLIN	BENNINGTON, VT.	TERT.	R	A				A			A			

nonmarine conditions of deposition. Some of the shales like those at Thedford, Ontario (1, 2), and Levis, Quebec (26), contain marine fossils at the locality where the samples were collected, and other formations are reported to contain marine fossils at localities other than where the shale was sampled. Both types of shale are listed in Table 1 as being deposited under the conditions of a marine environment. Type I and Type II hydrous micas are present in shales of marine as well as in those of nonmarine origin. Therefore, they are not helpful as an indicator of either environment of deposition. The shales of Pennsylvanian age that contain kaolinite, at New Glasgow and Woodburn, Nova Scotia (30, 31, 32), are considered by W. A. Bell (Personal communication, March 6, 1958) to be nonmarine in origin.

POTASH CLAY MINERALS IN CLAYS

Clay Minerals

The term clay is used here for a rock that lacks conspicuous lamination and contains abundant clay minerals. The clay minerals, which are hydrous silicates of aluminum, iron, and magnesium, impart plastic properties to the material especially when it is wet. A few clays, for example "flint clays," have little or

no plasticity when mixed with water. Clay is also used as a particle-size term for the material less than 2 microns, but some clays contain clay-mineral sheets that are larger than 2 microns. As a rock term clay is not used with the same connotation in all regions. In New England and Eastern Canada varved sediments are called varved clays. Although some of the sediments are distinctly laminated, the term varved clay is in general use and will be used here for this type of sediment. Some of the rock material used for brick making does not have the grade size of clay which in some classifications may include all the particles less than 4 or 5 microns. Geologists who mapped the clay deposits of Maine and New Hampshire (Trefethen *et al.*, 1947, p. 13; Goldthwait, 1951, p. 25; 1953, p. 7) used the term clay for the size of mineral grains regardless of composition. They classified grains smaller than .005 mm in diameter as clays. The results of 43 particle-size analyses of clayey samples made by E. Cormier in 1949 indicate that these clayey samples from Maine averaged 39 per cent clay, 37½ per cent silt, and 23½ per cent sand (Goldthwait, 1951, p. 25). In 1953 Goldthwait (p. 7) reported that the marine clays of New Hampshire, Maine, Massachusetts, New Brunswick, and the St. Lawrence Valley consist of clay, silt, and some sand. Considerable variation exists not only in the percentage of grain sizes from place to place but also in the percentage of clay minerals, which have a marked effect on the physical and burning properties of the clay. The clay at Saco, Maine (43), is of interest because it has high plasticity, abundant hydrous mica Type I, and was used in the manufacture of brick at one time according to reports of local citizens. The sample from Thomaston, Maine (42), contains a much higher proportion of grade sizes coarser than clay and a lower percentage of clay minerals than is found in the clay at Saco, Maine. Typical Quaternary clays of New England and Eastern Canada resemble the shales of the area in mineral composition; they are composed chiefly of potash clay minerals, or hydrous micas, and chlorite or vermiculite. Many clays contain Type I hydrous mica (Table 1, 7–9; 40–54), but others contain Type II hydrous mica (Table 1, 10–12; 14; 19–21; 24–25). The varved clay at Deschaillons, Quebec (25), is unusual in lacking chlorite or vermiculite and in having hydrous mica Type II as its only clay-mineral component.

Other Minerals

Minerals present in clays composed dominantly of potash clay minerals include the most varied mineral assemblages of any type collected from New England and Eastern Canada. Where the percentage of quartz larger than .005 mm approaches that in a silt the amounts of feldspars and other nonclay minerals increase proportionally. The feldspars observed include orthoclase, microcline, perthite, albite, and plagioclase. Fragments of unweathered green amphiboles are present in many samples. Zircon, epidote, zoisite, sphene, garnet, and tourmaline are minor constituents. The assemblages of minerals are similar to those observed by Allen (1930) during his petrographic studies of the weathered zones of glacial deposits in Illinois. Dolomite and calcite occur in shales and clays of marine and nonmarine origin. Also these minerals have been observed in thin sections filling veins cutting bedding and other depositional features. This indicates secondary deposition by carbonate solutions. Therefore, dolomite and calcite are not reliable indicators of the environments in which the clay minerals were deposited.

Conditions of Deposition

The clays of Quaternary age that are composed of potash clay minerals in this region were deposited under a variety of conditions. Some of them, like the silty clay at Riviere Cachee, north of Three Rivers, Quebec (24), contain marine microfossils and were deposited under marine conditions. Others are boulder clays containing some angular rock fragments that were deposited by glacial action (21). Many of them are varved clays that consist of a silty layer deposited during the summer and a fine-clay layer deposited during the winter. According to Antevs (1922) and others these annual layers or varves were deposited in fresh-water lakes that were supplied with sediments from the melting of glaciers during the Pleistocene. Hydrous micas Type I and Type II occur in both marine and nonmarine clays and therefore cannot be used as a guide to the environment in which the clays were deposited.

Kaolinite in Clays

X-ray Patterns

The kaolinite present in the clays collected during this investigation has the X-ray pattern

of well-crystallized kaolinite, and none has the "fire clay" type of pattern.

Sedimentary Kaolins

Clays in which kaolinite is the dominant clay mineral occur near Shubenacadie (37) and Middle Musquodoboit (34), Nova Scotia. Small amounts of hydrous mica Type I are present in samples from both localities, and minor amounts of montmorillonite are indicated in some samples from Shubenacadie. These moderately refractory sedimentary clays are used in the manufacture of pottery at Shubenacadie, Nova Scotia, and St. John, New Brunswick. The only fossils found in the clays at Shubenacadie and Middle Musquodoboit are meager plant remains, chiefly spores. The limited evidence suggests that these clays were deposited under nonmarine conditions.

In the early reports the age of these clays is given as Cretaceous (Ries, 1927, p. 591). J. Terasmae collected and studied the spore and pollen types present in a thin lignite seam in the clay at Shubenacadie and favors an Upper Jurassic age for the clay, mainly on the basis of the lack of dicotyledonous spores (Personal communication, March 24, 1958). A sample collected by Allen at Middle Musquodoboit was studied by Estella B. Leopold; it contained spores of dicotyledons. This suggested either a Cretaceous or younger age and introduced the possibility that the clay accumulated by reworking of sediments during the Quaternary. All the samples from Middle Musquodoboit lack feldspars, which are present in every clay of Quaternary age, from Eastern Canada and New England, that was studied. It is doubtful that reworking in the Quaternary could derive kaolinite from an older clay without contamination from feldspars that are present in the source rocks and in the clays deposited in this region during the Quaternary. Fragments of these mottled clays have been observed by Owen L. Hughes (Personal communication, October 17, 1955) in glacial till in this region. The close lithologic resemblance of the mottled clay at Middle Musquodoboit to the "mottled facies" of the Mattagami series of northern Canada described by McLearn (1927) has been observed by Hughes and others. The limited evidence does not permit a positive determination of the age of the clays at Shubenacadie and Middle Musquodoboit but favors a Late Jurassic or Early Cretaceous age.

Residual Kaolins

Samples of kaolin from Bennington, Vermont, at various depths from the surface to a depth of 155 feet were obtained through the courtesy of S. C. Lyons. These samples were taken during the joint drilling program by the Georgia Kaolin Corporation and Edgar Brothers in 1936, which established the presence of 6 million tons of clay in this deposit (Jacobs, 1940, p. 14).

Kaolinite, hydrous mica, and quartz are the principal minerals, and zircon and tourmaline are minor accessory minerals in the samples examined petrographically. No feldspars are present in the upper part of the deposits or in samples down to a depth of 155 feet in the drill holes studied. In one sample a broken fragment (6 by 12 mm) of milky quartz from a pegmatite or vein is present in the clay. This angular quartz has kaolinite etched on three surfaces and is too large to have been deposited by water with clay and silt-sized grains in a sedimentary clay deposit. Other angular milky quartz fragments were observed in the residue left after washing the clay that was mined and prepared for the market (Burt, 1928, p. 77). The samples of the clay available from the drilling project consisted of fragments too small for making thin sections. However, Burt (1928, p. 76) reported that thin sections of clay collected 15 feet from the surface at one mine contained kaolinite and kaolinized feldspars that were formed from orthoclase and albite. At another shallow deposit the clay rests on bedrock consisting of feldspathic gneiss that contains zircon and tourmaline similar to that in the overlying kaolinitic clay. He concluded that the clay is a residual deposit resulting from normal weathering of a Precambrian feldspathic gneiss during the Tertiary period. Considerable variation in vertical thickness of isolated clay deposits in a belt that has the same structural features and trend is characteristic of residual clay deposits. The quantity of hydrous mica and potash is less in sample 55 from a depth of 66 feet than that in sample 56 from a depth of 155 feet (Table 1). A decrease in potash or alkalies in the upper part of a clay deposit is also a characteristic feature of many residual deposits in which leaching has progressed downward from the surface to the lower part of the clay deposit.

Residual kaolin deposits at Cornwall, Connecticut, were described by Loughlin (1905). The shaft has caved and filled with water,

and samples could not be obtained from the old deposit. A sample of clay from West Cornwall was obtained through the courtesy of John B. Lucke; it is composed of kaolinite, angular quartz, and minor minerals.

Hydrothermal Kaolins

A deposit of kaolin associated with a zone of fracture in Precambrian quartzite occurs at St. Remi near Huberdeau in Amherst Township, Quebec. The white clay that was washed from this deposit has been used as a paper filler and in whiteware. The mine is no longer operating, and the shaft, which is more than 100 feet deep, is closed.

Osborne (1935, p. 69) considers that the quartzite was originally feldspathic and the feldspars were replaced by hydrothermal solutions coming from granites of the region. New quartz was formed in the quartzite in addition to the kaolin mineral and sericite and forms irregular stringers and pockets in the quartzite. The outer layer of some quartz crystals is clouded with the kaolin mineral, and this is interpreted as evidence that the formation of the kaolin mineral took place during the deposition of the secondary quartz.

Several samples of the kaolin (16, 17, 18) from St. Remi were examined to see if any dickite was present, because in the laboratory dickite is formed at a slightly higher temperature than that at which kaolinite crystallizes (Ewell and Insley, 1935). X-ray patterns and petrographic study revealed neither dickite nor the mosaic pattern of the clay minerals like that observed at Hobart Butte, Oregon, and other hydrothermal clay deposits (Allen, 1946, p. 133; Allen *et al.*, 1952, p. 6). Many books of micaceous plates of kaolinite are present. Hydrous mica, quartz, and a few rounded zircon grains are the principal minerals in this clay. This suggests that the replacement of the feldspars took place at a temperature below that at which dickite forms. The redistribution of the kaolinite and quartz and the known extension of the deposit to a depth exceeding 125 feet would be favored by Osborne's hypothesis of replacement by warm magmatic solutions rather than by solutions associated with normal weathering processes.

Mixed-Layer Clay Minerals in Shales and Clays

Mixed-layer clay minerals are those in which two clay minerals are interlayered. Mica and chlorite are the minerals that are interlayered and occur as minor constituents in the shales and clays of this region. These were recorded in the X-ray patterns of only 7 of the 56 samples that were studied. Four of the samples are from shales that contain hydrous mica as their chief clay mineral and chlorite. These shales were probably deposited under marine conditions. The clays with some mixed-layer clay minerals also contain abundant hydrous mica and chlorite; they belong to the Quaternary period, and one has been assigned to a marine environment and the other to a nonmarine environment. The evidence is not sufficient to draw definite conclusions as to the value of mixed-layer clay minerals as indicators of the conditions under which they were deposited.

Montmorillonite Minerals in Clays

Montmorillonite was recorded as a minor constitutent in only two samples collected during this investigation. One is from the nonmarine clays of Late Jurassic (?) age from Shubenacadie, Nova Scotia (37), and the other is from a silty clay at Riviere Cachee (24) near Three Rivers, Quebec, that contains Quaternary marine microfossils.

The montmorillonite-hydrous mica ratio of the clay fraction from shales has been suggested as a means of distinguishing marine from nonmarine formations. The Research Council of Alberta (1956, p. 24) reports that the montmorillonite-hydrous mica ratio for the nonmarine Oldman formation is somewhat higher than that for the marine Bearpaw formation of Late Cretaceous age in the same area. No other essential difference was found in the mineralogical composition of the samples collected across the contact of these formations in Alberta.

Montmorillonite is not present in sufficient amounts in the samples collected during this investigation to determine the montmorillonite-hydrous mica ratio or whether the origin of this montmorillonite is related to volcanic activity or reflects the source areas that contained the montmorillonite. The reader interested in the latter problem at other localities will find it discussed by Grim (1953, p. 356), Holmes and Hearn (1942, p. 71), Griffin and Johns (1958, p. 1574), Johns and Grim (1958, p. 186), and Weaver (1958, p. 839–861).

Clay Minerals in Relation to Geological Environments

General Considerations

An attempt has been made during this investigation to obtain detailed information on

the mineral composition and structure of clays and shales that would provide evidence of the geological environment in which the clay minerals were deposited. Over the years samples have been submitted to the writers by geologists and paleontologists with questions about the conditions of origin. The question that is most frequently asked is: were the clay minerals deposited under marine or fresh-water conditions? The evidence obtained during this investigation regarding the use of clays as indicators of depositional environments and a brief summary of current knowledge in this field is given on the following pages.

The statement appears in publications on engineering geology (Krynine and Judd, 1957, p. 133) that marine clays are more porous than fresh-water clays. This is explained as the result of sodium chloride in salt water acting as an electrolyte and causing the relatively remote colloidal particles as well as the close ones to flocculate and form large flocs and large voids. The hypothetical bond between the coarse silt particles and the chainlike particles of clay minerals surrounding large open voids is given by Casagrande (1932, p. 181). Both fresh-water and marine clays were studied under high magnifications, but no differences in structure of fresh-water and marine clays could be detected in thin sections.

The idea that abundant hydrous mica in a shale or clay indicates a marine environment is expressed from time to time; however, this inference is unwarranted. Potash clay minerals, or hydrous micas, are present in each of the 56 samples listed in Table 1 and range from a trace to abundant. In two samples (25, 27) they are the only clay minerals present. Sample 25 is from clay of nonmarine origin, and sample 27 is from a marine shale. A chloritic clay mineral occurs with the potash clay mineral in many of the samples from both marine and nonmarine beds. Therefore, this mineral is also of little value as an indicator of depositional environment.

Marine Clays

Hydrous mica and a chloritic mineral have been observed as constitutents of Recent marine sediments. Dietz (1941, Ph.D. thesis, Univ. Illinois, p. 68) suggested that hydrous mica develops largely by the alteration of montmorillonite from the source material carried to the sea. Murray and Harrison (1956, p. 368) report that cores from the Sigsbee Deep contain hydrous mica, montmorillonite, and chlorite, and in some cores montmorillonite is the most abundant clay mineral. Grim, Dietz, and Bradley (1949, p. 1806) found that a series of samples collected by the E. W. Scripps staff from the Gulf of California and the Pacific Ocean off the California coast contained hydrous mica, chloritic mica, montmorillonite, and kaolinite and that hydrous mica was the most abundant mineral in nearly all samples. They concluded that potassium is being taken up by the accumulating sediments mostly by "degraded illite," which is hydrous mica that reaches the site of deposition with a relatively low potassium content as a result of weathering processes, and that magnesium is being taken up by the accumulating sediments, probably in the formation of hydrous mica or chloritic micas. The widespread occurrence of montmorillonite in marine sediments is accepted as evidence that it is not lost quickly or completely in the diagenic processes, and its absence or scarcity in ancient marine sediments is caused by metamorphism rather than by diagenesis.

The investigations of Powers (1954), Grim and Johns (1954), and Johns and Grim (1958) indicate the presence of hydrous micas and chlorite in Recent marine sediments and the formation of these minerals from degraded micas and montmorillonite in a marine environment.

Grim and his colleagues (1949, p. 1806) observed that the kaolinite content of samples from the Pacific Ocean and Gulf of California was less than that in samples from the Colorado River. They concluded that kaolinite is slowly being lost in the marine sediments, probably by alteration to hydrous mica or chloritic mica. Another possibility is that kaolinite is concentrated by mechanical processes because of grain size and reaction to the flocculating agents in sea water, so that most of the large crystalline plates are deposited near shore or in the deltaic or fluviatile environments. Thus, only the fine poorly organized particles are scattered over the wide areas of the marine environment. Here these fine mixtures are diluted with abundant hydrous mica, give indistinct X-ray patterns, and are somewhat susceptible to change because of broken bonds and their fine particle size. Kaolinite has been known to survive contact with sea water during deposition, as is shown by the presence in the deltaic sediments of the Ione formation of California (Allen, 1929, p. 422) of marine fossils in contact with kaolinite and quartz

grains. In both the deltaic sediments of the Ione formation and the fluviatile sediments upstream kaolinite is the only clay mineral present, so that no apparent decrease or loss in kaolinite is evident. The presence of small amounts of kaolinite in Recent marine sediments is evidence that kaolinite has persisted in a modern marine environment but is not proof that large quantities of kaolinite reached the marine environment and were altered there to other clay minerals under the same conditions in which the limited quantity of relict kaolinite survived.

Nonmarine Clays

Sediments in which kaolinite is the dominant clay mineral were apparently deposited in a nonmarine environment. Some of these sediments like those of the Ione formation of California have sedimentary structures that suggest deposition in the deltaic and fluviatile environments (Allen, 1929). The Cretaceous kaolin deposits of Georgia and South Carolina were deposited in a deltaic or littoral environment according to Kesler (1957, p. 6). The underclays of Illinois (Grim and Allen, 1938, p. 1507) that contain abundant kaolinite were deposited in an aqueous environment that Weller (1930) considers to be nonmarine. His conclusion is supported by the presence of fresh-water limestones and abundant plant-root fossils, stigmariae, in many underclays.

K. M. Waagé (Personal communication) has found that a considerable decrease seaward in the proportion of kaolinite in the clay fraction at the shore line is a useful criterion in locating the boundary between contemporaneous nonmarine and marine parts of the Dakota sandstone of Cretaceous age in Colorado.

Grim and his colleagues emphasized (1949, p. 1806) that an abundance of kaolinite in an ancient sediment suggests that it was not deposited under marine conditions, and the results of our investigation are in keeping with this conclusion. The clays and shales in New England and Eastern Canada in which kaolinite is abundant (Table 1, 30, 31, 34, 37) lack marine fossils and contain plant remains and structures that have been interpreted as indicating deposition in a nonmarine environment.

The clay-mineral composition of varved clays from New England and Eastern Canada listed in Table 1 is similar to that in the clays from the same region that contain marine fossils and lack varves. No appreciable difference is evident in the mineral composition of clays that lack abundant kaolinite and were deposited during the Quaternary in fresh-water lakes and those deposited in the sea. The component clay minerals, hydrous micas and chlorite, were derived from similar source rocks and were deposited relatively rapidly without change in each environment. The source material has a greater effect on the mineral composition of these sediments than the environment in which deposition took place. The sharp lines of the X-ray patterns of hydrous mica Type I indicate well-crystallized plates of a grain size that gives a clear, distinct pattern. The similarity of the mineral composition of these clays to the shales of the same geographic region indicates a considerable contribution of these minerals from suitable source rocks. A significant portion of the varved clays probably came from shales, slates, phyllites, mica and chlorite schists, and mica-bearing gneisses, which are common in the glaciated areas. Rock flour ground from these rocks by glaciers no doubt was similar to the varved clays in mineral composition.

Clays Formed by Weathering and Hydrothermal Processes

The evidence for distinguishing clay deposits formed by hydrothermal processes from those formed by normal surface weathering has been reviewed by Grim (1953, p. 323–345) and by Kerr (1955, p. 531–540). The zonal arrangement characteristic of some hydrothermal deposits (Lovering, 1949, p. 65) has not been observed in the clay deposits of this region.

The preservation of relict textures of minerals and rocks in thin sections of clays from some localities (Allen, 1929, p. 383–388; 1930, p. 85; 1930, unpublished report, Illinois State Geological Survey; 1948, p. 622; Allen and Scheid, 1946, p. 303–309) has contributed evidence of the formation of the clay *in situ*. The samples of clay stored in jars for years by Lyons Bros. at Bennington, Vermont, and those collected at St. Remi, Quebec, were not satisfactory for making a series of thin sections that would record the changes in the deposits that had taken place from top to bottom. However, the minerals present and the observations of previous investigators indicate that the kaolin clays at Bennington are residual

clay deposits formed by weathering processes, and the kaolin at St. Remi is a replacement deposit formed by hydrothermal processes.

Conclusions

The potash clay minerals, or hydrous micas, and chlorite or vermiculite are the abundant clay minerals in the marine shales and clays, in nonmarine varved clays, and in glacial tills examined from this region. The presence of potash clay minerals and chlorite is not a reliable indication that the sediment containing them accumulated in a marine environment. Varved clays deposited in lakes have the same clay-mineral composition as clays containing Quaternary marine fossils and Paleozoic marine shales in the same area. Therefore the source area from which the clays were derived had a more pronounced influence on the clay-mineral composition of the resulting sediment than did the environment in which the clay minerals were deposited.

The clays and shales in which the fossils and structures indicate a nonmarine environment of deposition have abundant kaolinite. The kaolinite was eroded from a land area where kaolinite clays were present. Two types of kaolinite clays are known in this region. Residual kaolin was formed by surface weathering in New England probably during the Tertiary. Replacement kaolin was formed by hydrothermal solutions in Quebec during Precambrian time.

References Cited

Allen, Victor T., 1929, The Ione formation of California: Univ. Calif. Pub. Bull., Dept. Geol. Sci., v. 18, p. 347–448

—— 1930, Petrography of weathered zones of glacial deposits: Geol. Soc. America Bull., v. 41, p. 85–86

—— 1946, Sedimentary and volcanic processes in the formation of high alumina clay: Econ. Geology, v. 41, p. 124–138

—— 1948, Formation of bauxite from basaltic rocks of Oregon: Econ. Geology, v. 43, p. 619–626

Allen, Victor T., and Scheid, Vernon E., 1946, Nontronite in the Columbia River region: Am. Mineralogist, v. 31, p. 294–312

Allen, Victor T., Loofbourow, J. S., and Nichols, R. F., 1952, The Hobart Butte high alumina clay deposit, Lane County, Oregon: U. S. Geol. Survey Circ. 143, p. 1–11

Antevs, Ernst, 1922, The recession of the last ice sheet in New England: Am. Geog. Soc. Research Ser., No. 11, 120 p.

Burt, Frederick A., 1928, Origin of the Bennington kaolins: State Geologist Vt. Rept., 1927–1928, p. 65–85

Casagrande, Arthur, 1932, The structure of clay and its importance in foundation engineering: Boston Soc. Civil Engineers, v. 19, p. 168–209

Ewell, R. H., and Insley, H., 1935, Hydrothermal synthesis of kaolinite, dickite, beidellite, and nontronite: U. S. Natl. Bur. Standards Jour. Research, v. 15, p. 173–186

Goldthwait, Lawrence, 1951, The glacial-marine clays of the Portland-Sebago Lake region, Maine: Maine Geol. Survey Rept. State Geologist, 1949–1950. p. 24–34

—— 1953, Preliminary report on clay deposits of southeastern New Hampshire, pt. XV. Mineral resource survey: N. H. State Plan. and Devel. Comm., 14 p.

Griffin, J. J., and Johns, William D., 1958, Clay-mineral composition of sediments from Mississippi River and major tributaries (Abstract): Geol. Soc. America Bull., v. 69, p. 1574

Grim, Ralph E., 1953, Clay mineralogy: N. Y., McGraw-Hill Book Co., Inc., 384 p.

Grim, Ralph E., and Allen, Victor T., 1938, Petrology of the Pennsylvanian underclays of Illinois: Geol. Soc. America Bull., v. 49, p. 1485–1514

Grim, R. E., Dietz, R. S., and Bradley, W. F., 1949, Clay mineral composition of some sediments from the Pacific Ocean off the California coast and the Gulf of California: Geol. Soc. America Bull., v. 60, p. 1785–1808

Grim, Ralph E., and Johns, William D., 1954, Clay mineral investigation of sediments in the northern Gulf of Mexico, p. 81–103 *in* Swineford, A., and Plummer, N. V., *Editors*, Clays and clay minerals: Natl. Research Council Pub. 327, 498 p.

Holmes, R. S., and Hearn, W. E., 1942, Chemical and physical properties of some of the important alluvial soils of the Mississippi Drainage Basin: U. S. Dept. Agriculture Tech. Bull. 833, 81 p.

Jacobs, E. C., 1940, The clay deposits and clay industries of Vermont: State Geologist Vt. Rept., 1939–1940, p. 14–16

Johns, William D., and Grim, Ralph E., 1958, Clay mineral composition of Recent sediments from the Mississippi River delta: Jour. Sed. Petrology, v. 28, p. 186–199

Kerr, Paul F., 1955, Hydrothermal alteration and weathering, p. 525–544 *in* Poldervaart, A., *Editor*, Crust of the earth: Geol. Soc. America Special Paper 62, 762 p.

Kesler, Thomas L., 1957, Environment and origin of the Cretaceous kaolin deposits of Georgia and South Carolina: Ga. Geol. Survey Min. Newsletter, v. 10, No. 1, p. 1–8

Krynine, D. P., and Judd, W. R., 1957, Principles of engineering geology and geotechnics: N. Y., McGraw-Hill Book Co., Inc., 730 p.

Loughlin, Gerald F., 1905, The clays and clay industries of Connecticut: Conn. Geol. Survey Bull. 4, 121 p.

Lovering, T. S., 1949, Rock alteration as a guide to ore-East Tintic district, Utah: Econ. Geol., Mon. 1, 65 p.

McLearn, F. H., 1927, Mattagami series: Geol. Survey Canada Summary Rept. 1926, Pt. C, p. 20–27

Murray, H. H., and Harrison, J. L., 1956, Clay mineral composition of Recent sediments from

Sigsbee Deep: Jour. Sed. Petrology, v. 26, No. 4, p. 363–368

Osborne, F. Fitz, 1935, Sainte-Agathe-Sainte Jovite map area: Ann. Rept. Quebec Bur. Mines, Pt. C, p. 69

Osborne, F. Fitz, 1954, The petrology of the Charny formation: Geol. Assoc. Canada Proc., v. 6, pt. 2, p. 111–114

Powers, Maurice C., 1954, Clay diagenesis in the Chesapeake Bay area, p. 68–80 *in* Swineford, A., and Plummer, N. V., *Editors*, Clays and clay minerals: Natl. Research Council Pub. 327, 498 p.

Research Council of Alberta, 1956, Clay mineralogy: Rept. 76, 37th Ann. Rept., p. 1–48

Ries, Heinrich, 1927, Clays; their occurrence, properties, and uses: N. Y., John Wiley & Sons, Inc., 613 p.

Trefethen, J. M., Allen, H., Leavitt, L., Miller, R. N., 1947, Preliminary report on Maine clays: Maine Dev. Comm. State Geologist Rept. 1945–1946, p. 10–46

Weaver, Charles Edward, 1958, Effects and geologic significance of potassium "fixation" by expandable clay minerals derived from muscovite, biotite, chlorite, and volcanic material: Am. Mineralogist, v. 43, p. 839–861

Weller, J. M., 1930, Cyclical sedimentation of the Pennsylvanian period and its significance: Jour. Geology, v. 38, p. 97–135

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Short Notes

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MAGNETIC SUSCEPTIBILITY AND FUSION DATA FOR SOME VOLCANIC ROCKS FROM SOUTHWESTERN NEW MEXICO

By Joseph G. Wargo

This note presents the results of studies made on (1) the magnetic susceptibilities of some rhyolitic rocks and (2) refractive indices of fused rhyolitic rocks, from southwestern New Mexico.

Figure 1.—Location Map

The volcanic rocks were collected from several unnamed volcanic formations found in a 7½-minute quadrangle bounded by Lats. 32° 45′ and 32° 52′ 30″ N. and Longs. 108° 30′ and 108° 37′ 30″ W., in the northern Big Burro Mountains, Grant County, New Mexico (Fig. 1). Elsewhere, the writer (Wargo, 1958) has referred to this area as the "Schoolhouse Mountain area" and has briefly described the volcanic stratigraphy and structure. The stratigraphic section in the volcanic rocks consists of three major units: (1) a thin lower andesite unit, (2) a thick medial rhyolite unit, and (3) a thin upper basalt unit. The medial rhyolite unit is the most extensive and consists mostly of rhyolitic pyroclastic rocks, flows, and welded tuffs. Similar and possibly genetically and temporally related rocks are found throughout southwestern New Mexico (Wargo, 1959). The data presented here refer to certain well-defined beds within the medial rhyolite unit.

Table 1.—Composition of some Rhyolitic Rocks from Southwestern New Mexico

Doris Thaemlitz, analyst

	1*	2	3	4	5*
SiO_2	71.53	72.68	75.45	72.09	72.12
Al_2O_3	..	14.28	13.02	14.18	..
Fe_2O_3	1.35	1.34	.85	1.65	.83
FeO	.23	.26	.14	.23	.25
MgO	.50	.32	.21	.36	.16
CaO	.98	.51	.34	1.14	.63
Na_2O	2.96	3.50	3.82	3.94	3.94
K_2O	4.37	5.02	4.82	4.53	4.34
H_2O+	..	.79	.30	.47	..
H_2O-	..	.29	.35	.51	..
TiO_2	..	.26	.18	.31	..
P_2O_5	..	.07	.03	.09	..
MnO	..	.05	.05	.06	..
Total	81.92	99.37	99.56	99.56	82.27

1. Vitric crystal tuff, sec. 6, T. 17 S., R. 16 W.
2. Brown tuff breccia, sec. 6, T. 17 S., R. 16 W.
3. Gray rhyolite, sec. 31, T. 16 S., R. 16 W.; sec. 5, T. 17 S., R. 16 W.; sec. 15, 16, T. 17 S., R. 17 W.
4. Reddish welded tuff, sec. 3, 4, 9, T. 17 S., R. 17 W.
5. Vitrophyre, sec. 8, T. 17 S., R. 16 W.; sec. 22, T. 17 S., R. 17 W.

* Partial analysis

In order to provide material for laboratory determination of chemical composition and refractive index of fused rock, numerous samples were collected in the field, after a detailed geo-

logic map had been made. The sample traverses were of two kinds: (1) those having broadly spaced stations (up to a quarter of a mile between sample points), and (2) those having closely spaced stations (200 feet between sample points). The sampling therefore provides material for rather detailed studies of certain beds as well as for studies of the grosser aspects of the whole formation. At each sample station, about 8 pounds of rock was collected from within a circle having a 50-foot radius. Only fresh, unaltered, and unweathered rocks were collected.

Five rock types were sampled: (1) soft, white vitric crystal tuff; (2) hard, brown tuff breccia consisting of reddish and brownish fine-grained lithic fragments cemented by a brownish tuff; (3) dense, hard, gray rhyolite flow containing streaky vesicles lined with quartz; (4) hard, reddish welded tuff showing numerous examples of flattening and squeezing of glass shards; and (5) hard, black, shiny vitrophyre.

Composite samples were made of each rock type. The samples were analyzed at the Rock Analysis Laboratory of the University of Minnesota. The results of these chemical analyses are given in Table 1. In addition numerous SiO_2 analyses were made from samples taken from all five rock types.

Magnetic susceptibilities of the various kinds of rhyolitic rocks were determined on the outcrop, using a magnetic susceptibility meter supplied by the Bear Creek Mining Company. Essentially, the instrument consists of three coils, a battery, and a bridge circuit. Measurement of susceptibility is made indirectly by determining the induction in the coils when they are on a tripod at some distance from the outcrop and again when they are properly placed on the outcrop. The difference in inductance, which is due in part to the presence of magnetic materials in the rock, is calibrated in terms of susceptibility, and the susceptibility values are read from a calibration chart. A complete analysis of the theoretical aspects

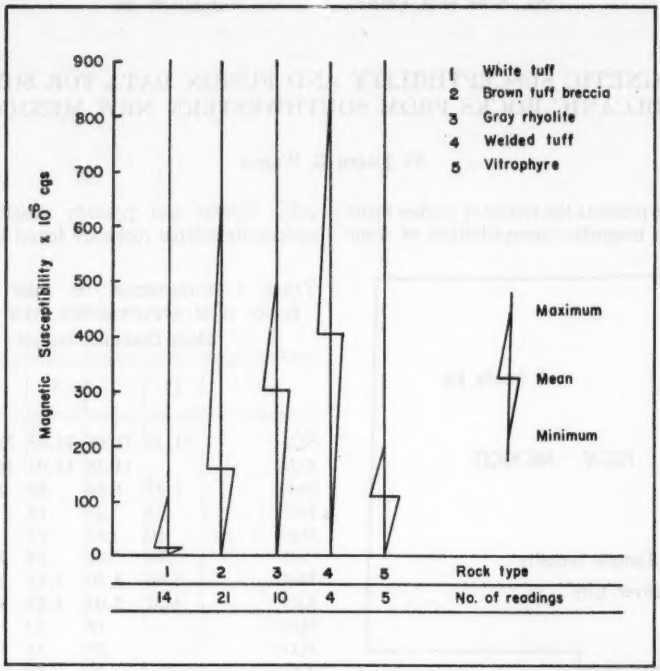

FIGURE 2.—MAXIMUM, MINIMUM AND MEAN MAGNETIC SUSCEPTIBILITIES OF SOME RHYOLITIC ROCKS FROM SOUTHWESTERN NEW MEXICO

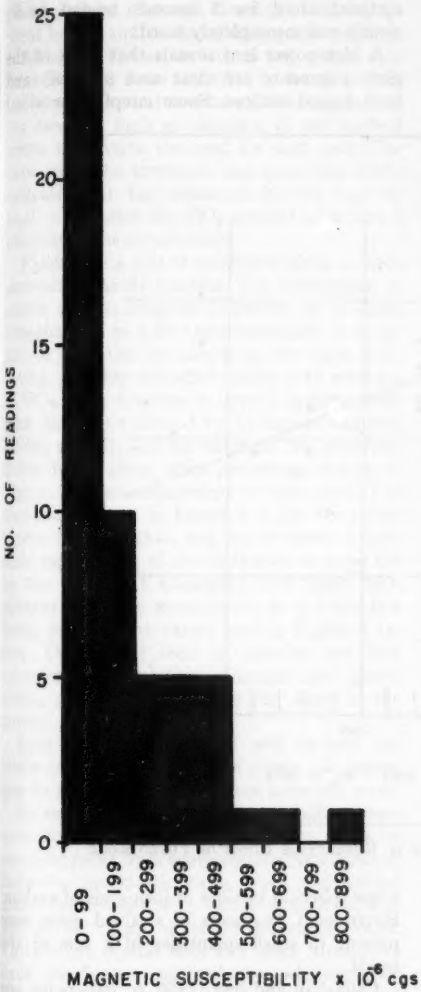

FIGURE 3.—HISTOGRAM OF MAGNETIC SUSCEPTIBILITY READINGS ON RHYOLITIC ROCKS FROM SOUTHWESTERN NEW MEXICO

of the instrument, as well as a description of the circuitry, is given by Mooney (1952).

The average, maximum, and minimum values of magnetic susceptibility for each rock type are shown in Figure 2. The rhyolitic white tuffs have the lowest average susceptibility and the least dispersion of values. Vitrophyre also has a low average susceptibility. The brown tuff breccia, gray rhyolite, and red welded tuff show a considerable dispersion of magnetic susceptibility values and have average susceptibilities more than 2 to 4 times that of the white tuff. A histogram of the magnetic susceptibilities for all the rocks is shown in Figure 3. The plot shows that a distribution curve would be skewed to the left and that in general the magnetic susceptibilities for these rhyolitic rocks tend to be low.

The process of fusing powdered volcanic rock to a glass and the determining of the refractive index of the glass is termed the fusion method. The relationship between the chemical composition of a glass and its refractive index has been noted by George (1924) and Mathews (1951), among others. George (1924) studied the chemical compositions and refractive indices of natural glasses and obtained a correlation between these variables. Mathews (1951) made chemical analyses of rock, fused the rock and determined the index of the glass, and gave a comprehensive discussion of the relationships between composition and refractive index. Callaghan and Sun (1956) used Mathews' methods to study rocks of various compositions from New Mexico and developed curves showing the relationship between SiO_2, FeO, Fe_2O_3, CaO, and MgO content and the refractive index of the fused rock. In order to develop additional data for the rhyolitic flows and pyroclastic rocks, the writer studied the refractive index of fused specimens from the area described above.

The following steps were taken, in the preparation of samples for fusion: (1) collection of the samples; (2) crushing to about minus 10 mesh or smaller; (3) splitting the sample to obtain a fraction for chemical analysis; the reject from this split was used for fusion study; (4) splitting the reject from (3) and grinding to form minus 80 mesh and minus 200 mesh fractions; and (5) splitting the minus 80 mesh and minus 200 mesh fractions to form 100-gram samples for fusion and reference.

The heat for the fusing of the powdered samples was provided by a Lincoln arc torch holding carbon electrodes. A Lincoln "Idealarc 180" welding unit operating at 160 amps, 25 volts supplied the current. About 0.5 gram of powdered rock was placed on a 6- by 1½- by 1¼-inch copper block, which rested in a pan of water. The arc was struck over the powder, causing some of the powder to melt. The high conductivity of the copper allowed rapid heat escape, so that the bead cooled rapidly to a glass. At no time was the arc brought so close to the copper that melting of the block could

occur, and an examination of the block after each run revealed no sign of pitting. The fact that the bead formed in a bed of rock powder further reduced the chances for copper contamination.

material arced for 3 seconds tended to be cloudy and incompletely fused.

A high-power lens reveals that most of the glass fragments are clear and isotropic and have jagged outlines. Some samples contained

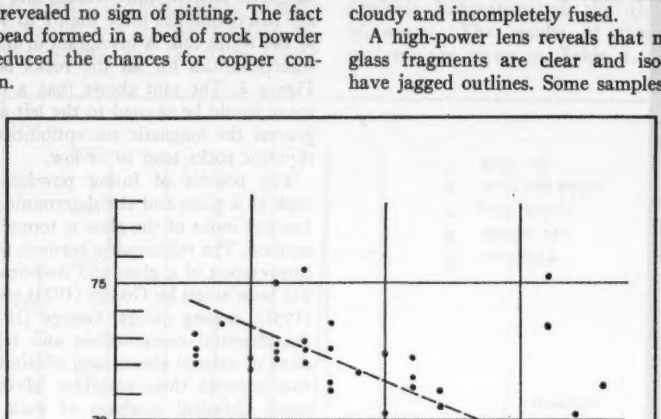

FIGURE 4.—PLOT OF SILICA PERCENTAGE *vs.* REFRACTIVE INDEX OF FUSED ROCK

The arc was maintained for the following lengths of time, for each sample: (1) minus 80 mesh fraction, 3 seconds; (2) minus 200 mesh fraction, 3 seconds; and (3) minus 200 mesh fraction, 6 seconds.

The glass beads formed were crushed in a ceramic mortar to about minus 150 mesh and examined with a petrographic microscope, using a high-power lens. Refractive indices were determined by the use of oils, at a temperature of 25° C. The light source was a Shannon Lab-arc which provided monochromatic sodium light. Only the beads from the minus 200 mesh powder arced for 6 seconds are considered here. No significant difference could be found between the minus 200-mesh, 6-second beads and the minus 200-mesh, 3-second beads. The beads made from the minus 80-mesh a few spherical bubbles or inclusions of carbon. Birefringent fragments of strained glass were present in small quantities in a few of the beads.

Callaghan and Sun (1956, p. 766) point out that the composition of the bead is not necessarily the same as the composition of the powder from which it was made. This may be due to incomplete fusion of certain constituents or to loss of some of the more volatile constituents during fusion. The effect of contamination of the bead by the atmosphere has not been studied but is thought to be small. The small size of the final sample may contribute a random error to the process, although fine grinding tends to reduce the error. Errors due to contamination during sample preparation have been minimized by the use of ceramic grinding

plates and by thorough cleaning of the equipment before and during crushing and splitting.

Although the errors mentioned above may be reduced by more attention to detail, such as fusion in an inert atmosphere and finer grinding, the inherent limit of accuracy of the method seems to obviate the need for such care. The data from the literature and from this study indicate that the refractive indices may be used to estimate the SiO_2 content to within 3 per cent of the actual value.

Figure 4 is a plot of refractive index *vs.* SiO_2 percentage for 37 samples. The distribution of points in this diagram is related by a single straight line, as a first approximation. It is apparent that the line slopes to the right, indicating a higher refractive index with decrease of SiO_2. This direction of slope is in accordance with the curve plotted by Callaghan and Sun (1956, p. 762), and the values of the refractive index for a given silica percentage shown in Figure 4 are also equivalent to their curve. The amount of slope in Figure 4 is less than that shown by Callaghan and Sun's curve. A possible explanation of this difference in slope lies in the fact that Callaghan and Sun's SiO_2 content data are recalculated to a water-free basis, whereas the values used in Figure 4 are not. This would tend to increase the SiO_2 amounts in the highly siliceous and glassy rocks, and perhaps steepen the slope of the curve.

Four or five points fall well beyond the limits of the central group of points. No reason can be given for this departure from the norm.

In summary, magnetic susceptibility measurements have been made on five types of rhyolitic rocks from southwestern New Mexico, including pyroclastic rocks, flows, and welded tuffs. The susceptibility values range from 0 to 825×10^{-6} c.g.s. units. In general, rhyolitic white vitric crystal tuffs and black vitrophyre have the lowest susceptibility, whereas gray rhyolite flows, brown tuff breccias, and red welded tuffs have somewhat higher susceptibility values. A histogram combining data for all rock types studied suggests that a low susceptibility is common for these rhyolitic rocks.

Fusion studies made on the same rhyolitic rocks substantiate data derived by other workers and add to the information on the relationship between refractive index and SiO_2 content of these rocks. In general, the refractive index of the fused rock increases as the SiO_2 content decreases. The arc-welder technique described here appears to hold promise in the preparation of glass beads for such studies.

The writer is indebted to the Rocky Mountain District of the Bear Creek Mining Company, a wholly owned subsidiary of the Kennecott Copper Corporation, for subsidizing portions of this study and for permitting publication of the results. Thanks are due Mr. Franklin Clark and Mr. John Russell of the Bear Creek Mining Company for reading the manuscript. Dr. Ming-Shan Sun of the New Mexico Bureau of Mines and Mineral Resources offered several helpful suggestions. The glass beads used in the fusion study were made by Mr. John Russell.

References Cited

Callaghan, Eugene, and Sun, Ming-Shan, 1956, Correlation of some igneous rocks of New Mexico by the fusion method: Am. Geophys. Union Trans., v. 37, p. 761–766

George, W. O., 1924, The relation of physical properties of natural glasses to their chemical composition: Jour. Geology, v. 32, p. 353–372

Mathews, W. H., 1951, A useful method for determining the approximate composition of fine grained igneous rocks: Am. Mineralogist, v. 36, p. 92–101

Mooney, H. E., 1952, Magnetic susceptibility measurements in Minnesota, Part I: Geophysics, v. 17, p. 531–543

Wargo, J. G., 1958, Structure and volcanic stratigraphy in the Schoolhouse Mountain area, Grant County, New Mexico (Abstract): Geol. Soc. America Bull. v. 69, p. 1748

—— 1959, Sequence of volcanic rocks in southwestern New Mexico (Abstract): Geol. Soc. America Bull., v. 70, p. 1754

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BOULDER IN VARVED CLAY AT STEEP ROCK LAKE, ONTARIO, CANADA

BY R. M. HARDY AND R. F. LEGGET

Varved glacial clays of unusual extent have been revealed by the unwatering of Steep Rock Lake, in western Ontario, an operation carried out in order to gain access to the high-grade iron ore beneath the bed of the lake. Samuel (1945) and Legget (1958) have described the initial development and methods of operation. Pearson (1957) has published an account of the final phase of the unwatering operation with which the first author has been associated as a consultant. A detailed study of typical exposures of the varved clay was made by Antevs (1951). Laboratory studies have been carried out primarily for engineering purposes (Legget and Bartley, 1953). Details of more detailed laboratory studies of these varved soils have been given by Eden (1955).

Legget and Bartley (1953) and Eden (1955) have raised questions about the process of sedimentation that resulted in the varves as now exposed by the excavation. Similar questions are raised by the appearance of the sediments adjacent to a small boulder exposed during dredging operations and photographed by Hardy (Pl. 1). The boulder was found near the center of Falls Bay, which is in the northeast corner of the lake and includes about 7 square miles of water surface. Its location was approximately 2000 feet from the northerly shore line, at an elevation of about 912 feet. When dewatering started in 1944 the lake level was 1263 feet, the lake-bed clay surface near the boulder was about 1000, and the bedrock elevation was about 825. Antevs (1951) suggests that the water level of "Lake Johnston", the glacial stage of the modern lake, stood at 1363 feet.

The freshly dredged clay surfaces are treacherous, and it was not, therefore, possible to recover the boulder for examination. It was, however, about 8 inches in equivalent diameter, and the photograph shows it to be partially rounded. It is the only deposition of its kind which either of the authors has seen or heard of during the course of their many visits to Steep Rock Lake.

It may be assumed that the boulder was dropped from floating ice in the lake waters above. If the maximum level suggested by Antevs be assumed, the boulder fell through a depth of some 450 feet of water before coming to rest on the clay lake bed, but its velocity of descent is indeterminate because ice may have been adhering to it. Careful examination of the photograph, however, shows that it landed on a dark layer and pierced this layer as well as the immediately underlying light-colored layer, but was held by the second dark layer which was only about 2 inches below the lake-bed surface. This is confirmed by the "curling" of the two layers to the lower right of the boulder. It follows, therefore, that the surface layers of the clay onto which the boulder fell, even though then so recently deposited, must have possessed appreciable shearing strength. Correspondingly, this gain in shear strength must have been rapid, much more so than is usually encountered in normally deposited soils.

The weight of the boulder compressed or consolidated the soil for several inches below the surface upon which it came to rest. It is therefore surprising to note that any subsequent consolidation of these same layers because of the weight of about 100 feet of overburden was not so great as that produced locally by the submerged weight of this relatively small boulder. It follows that there must have been a substantial increase in particle bond strength, and therefore a reduction in compressibility, in the interval between the time of the fall of the boulder and the completion of the process of sedimentation up to the final level of the bottom of the lake.

This suggestion is confirmed by the clearly defined steep slope of the first varve to be deposited on top of the boulder. The maximum slope of this exceeds 45° so that it is in striking contrast to the degree of slope normally encountered in such subaqueous deposits. Such a rapid build up of thixotropic forces between soil particles could account for other phenomena of the Steep Rock soils, already reported, such as the natural moisture contents higher than the liquid limits, and a high degree of sensitivity.

References Cited

Antevs, Ernst, 1951, Glacial clays in Steep Rock Lake, Ontario, Canada: Geol. Soc. Amer. Bull., v. 62, p. 1223–1262

Eden, W. J, 1955, A laboratory study of varved clay from Steep Rock Lake, Ontario: Am. Jour. Sci., v. 253, p. 659–674

Legget, R. F., 1958, Soil engineering at Steep Rock Iron Mines, Ontario, Canada: Inst. Civ. Eng. Proc., v. 11, p. 169–188; Discussion, v. 13, p. 93–117

Legget, R. F., and Bartley, M. W., 1953, An engineering study of glacial deposits at Steep Rock Lake, Ontario, Canada: Econ. Geology, v. 48, p. 513–540

Pearson, P. D., 1957, The Caland Project at Steep Rock Lake: Eng. Jour., v. 40, p. 1657–1664

Samuel, W., 1945, Steep Rock iron mines; Discovery and development: Eng. Jour., v. 28, p. 4–9

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BOULDER EMBEDDED IN VARVED CLAY IN THE BED OF STEEP ROCK LAKE, ONTARIO, CANADA

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SYNGENETIC BLEACHED BORDERS ON DETRITAL RED BEDS OF THE FOUNTAIN FORMATION, FRONT RANGE, COLORADO

By John F. Hubert

The Permo-Pennsylvanian Fountain formation, one of the well-known red beds of the Western United States, crops out extensively along the eastern flank of the Front Range in Colorado. The Fountain formation, excluding the basal Glen Eyrie shale member, is a continental deposit, consisting of alternating lenses of stream-channel, stream flood-plain, and deltaic-lacustrine sediments (J. F. Hubert, 1958, Ph.D. Thesis, Pa. State Univ.). Between Eldorado Springs and Colorado Springs, the Fountain formation averages 80 per cent channel arkoses, 17 per cent flood-plain micaceous arkoses, and 3 per cent deltaic-lacustrine quartzose arkoses. The thickness of the Fountain formation ranges from 4526 feet at Colorado Springs to 1119 feet at Eldorado Springs.

The rocks in the Fountain formation average 71 per cent red, 28 per cent white, and 1 per cent green beds. The red color is due to detrital hematite which stains the grains and clayey matrix. The white color occurs commonly as bleached borders on red, trough cross-laminated conglomerates and sandstones (channel deposits) adjacent to deep-red, clayey sandstones and siltstones (flood-plain sediments). Much of the white arkose occurs as these bleached white borders, except for primary white arkose beds in the upper third of the Fountain. Bleached white borders are abundant throughout the lower two-thirds of the Fountain formation but are less common in the upper one-third, where they are developed on about one-quarter of the red arkose beds.

The bleached white borders average between 6 inches and 2 feet in thickness and occur on the top and bottom of a red arkose[1] bed between two redstone[2] beds (Pl. 1, fig. 1). The contacts between the white borders of the red arkose beds and the redstone beds are abrupt. The bleached white borders are transitional with the red central portions of the arkose beds; irregularly shaped, residual red patches occur in many of the transitional zones between the white borders and the red central portions of the arkose beds (Pl. 1, fig. 2). Thick (20–30 feet) redstone beds are usually accompanied by thicker white borders on the adjacent red arkose beds than are the thin (several inches) redstone beds. Very thin (a few inches) anastomosing redstone beds, spaced less than about 1 foot apart vertically, however, are commonly interbedded throughout with bleached white arkose.

The channel arkoses and flood-plain redstones are lenses; each is commonly composed of a number of beds. The lenses are laterally interstratified to form the alternation of the arkose and redstone rock types observed at each outcrop. Where the ends of redstone lenses are exposed, a bleached white border on the adjacent red arkose completely encloses the redstone in a three-dimensional envelope (Pl. 1, fig. 1). The white borders are not the result of deposition of nonhematite-stained detritus, because both the lower and upper borders of the arkose beds are white, and individual cross-laminae continue across the red–white contacts. Because the bleached white borders are restricted to envelopes on the red arkose beds, they are clearly not the result of recent weathering or mildly reducing groundwater solutions. Therefore, these bleached white envelopes suggest that the change in color from red to white is an early reaction that

[1] The term arkose here denotes the relatively clay-free channel arkoses which are texturally conglomerates and sandstones. In 16 random arkose samples, the proportion of matrix (detritus less than 0.03 mm in size) averages 9 per cent and is consistently less than 17 per cent. Authigenic cements, largely calcite and K feldspar, range up to 20 per cent and average 4 per cent. The ratio of Q (quartz plus chert plus metaquartzite rock fragments): F (feldspar plus granitic rock fragments): M (mica plus micaceous metamorphic rock fragments) among the grains (detritus larger than 0.03 mm) in the arkoses is 44:53:3.

[2] Redstone denotes the flood-plain micaceous arkoses which are deep-red, clayey sandstones and siltstones. The redstones in a few places contain oxidized plant remains and fossil soil horizons. The proportion of matrix in 16 random redstone samples averages 35 per cent and is consistently not less than 19 per cent. Authigenic cements do not exceed 1 per cent in any of the samples. The Q:F:M ratio among the grains in the redstones is 41:43:16. Krynine (1950, p. 103) used the term redstone for flood-plain sediments of similar mineralogy and texture in the red, arkosic rocks of the Triassic of Connecticut.

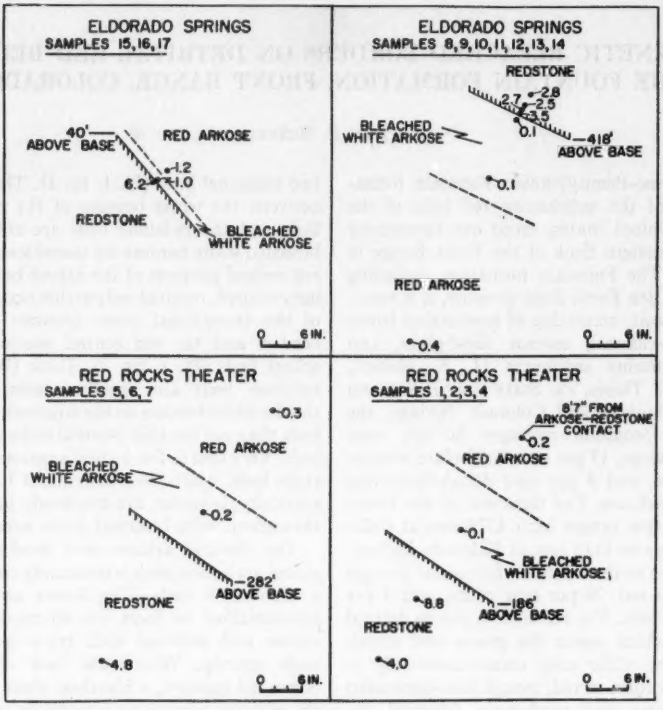

FIGURE 1.—IRON CONTENT OF FOUNTAIN SAMPLES

Outcrop sketches show locations of samples. The numbers refer to the percentage of iron in each sample. *Cf.* Table 1.

probably occurred during compaction of the beds with concomitant expulsion of connate water.

In a few localities a redstone cobble in the Fountain arkose conglomerates is completely enveloped by a bleached white border on the adjacent red arkose. The redstone cobbles clearly localized syngenetic bleaching of the red arkoses. Decaying plant remains within the redstone cobbles may have initiated bleaching of the surrounding red arkose, as discussed below.

It was necessary to determine which of the following two reactions was involved in bleaching of the red arkose beds: (1) Ferric iron may have been simply reduced to ferrous iron without loss of iron, under which condition the total elemental iron content in the white borders would remain the same as that in the red central portion of each arkose bed. (2) Ferrous

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PLATE 1.—SYNGENETIC BLEACHED WHITE BORDERS ON RED ARKOSE BEDS

FIGURE 1.—Stream-channel arkose and flood-plain redstone (dark-red rock beneath hammer) beds about 40 feet above base of Fountain formation at Eldorado Springs, Colorado. Top of beds is at right side of photograph. Syngenetic bleached white borders on the red arkoses completely envelope both the deep-red, clayey redstone beds.

FIGURE 2.—Flood-plain redstone (dark bed at top of specimen) adjacent to bleached white border on red arkose bed. Residual red patches occur in the white border of the arkose. Specimen was collected from block shown in Figure 1. Photograph shows cut surface of rock. Samples 15, 16, and 17 of Table 1 and Figure 1 are from this specimen.

FIGURE 1

FIGURE 2

SYNGENETIC BLEACHED WHITE BORDERS ON RED ARKOSE BEDS

iron from the borders of the red arkose beds may have been removed in solution, and ferric iron precipitated either in the adjacent redstones or carried away in solution from the arkose–redstone contact. To test these two hypotheses the amount of elemental iron was determined by X-ray fluorescence in samples of redstone, bleached white arkose border, and central red arkose collected at each of four localities (Table 1; Fig. 1).

The total elemental iron content of each sample was determined by X-ray fluorescence. Each sample was ground to pass a U. S. Standard mesh number 230 (0.006 mm) in a mullite mortar and pestle to avoid iron contamination. H. S. McKinstry, of the Mineral Constitution Laboratory of The Pennsylvania State University, kindly furnished mixed iron and silica samples for internal standards. All samples were run consecutively to avoid daily variation in the crystal analyzer. The quantitative iron determinations are subject to error in the assumption that the Fountain samples are composed solely of iron and silica as are the standards, and because of nonuniformity during sample preparation in packing, orientation, and particle size of the ground powder. Although the magnitude of these errors is unknown, the relative iron contents of the samples are established by the counts per second. The counts per second are an average of four readings, each reading based on a 40-second counting interval.

The proportions of elemental iron in the Fountain samples (Table 1; Fig. 1) lead to two conclusions. The total elemental iron content of the white arkose borders is consistently lower than the central portion of the red arkoses at each location, indicating that iron has been removed from the white borders. Iron has been concentrated in the redstones (Table 1; Fig. 1, samples 2 and 11) adjacent to the white borders of the red arkose beds in a narrow zone a few inches thick.

These results agree with those of Keller (1929) who found that bleached spots in red sediments are not completely explained by reduction of ferric iron to ferrous iron within the bleached portion, but involve removal of iron from the bleached spots. Keller suggested that hydrogen sulfide, either from decaying organic matter, sulfate-reducing bacteria, or sulfide waters, first reduced the relatively insoluble ferric iron to ferrous iron, which then was carried out of the bleached areas in solution in carbonate water.

TABLE 1.—IRON CONTENT OF FOUNTAIN SAMPLES BY X-RAY FLUORESCENCE

Outcrop	Distance of arkose-redstone contact from base of Fountain (Feet)	Sample No.	Rock type	Lithology	Distance from contact* (Inches)	Counts per second	Per cent Fe
Eldorado Springs	40	17	Red arkose	Pebbly sandstone	+2.0	1821	1.2
		16	White arkose	Pebbly sandstone	+0.2	1711	1.0
		15	Redstone	Clayey sandstone	−0.2	4612	6.2
Eldorado Springs	418	14	Redstone	Clayey siltstone	+3.3	2760	2.8
		13	Redstone	Clayey siltstone	+1.6	2611	2.5
		12	Redstone	Sandy, clayey siltstone	+1.0	2725	2.7
		11	Redstone	Sandy, clayey siltstone	+0.1	3164	3.5
		10	White arkose	Coarse sandstone	−0.1	81	0.1
		9	White arkose	Pebbly sandstone	−9.6	101	0.1
		8	Red arkose	Pebbly sandstone	−36.0	663	0.4
Red Rocks Theater	282	7	Red arkose	Coarse sandstone	+28.0	506	0.3
		6	White arkose	Fine conglomerate	+6.0	449	0.2
		5	Redstone	Clayey sandstone	−24.0	3835	4.8
Red Rocks Theater	186	4	Red arkose	Coarse sandstone	+103.0	313	0.2
		3	White arkose	Coarse sandstone	+7.0	101	0.1
		2	Redstone	Clayey siltstone	−5.0	6213	8.8
		1	Redstone	Sandy, clayey siltstone	−17.0	3426	4.0

* + Indicates sample above arkose–redstone contact
− Indicates sample below arkose–redstone contact

The following sequence is suggested to explain the origin of the syngenetic bleached white borders on the Fountain red arkose beds: (1) Ferric iron in the channel arkoses was reduced to ferrous iron by hydrogen sulfide, which perhaps originated from decaying plants[3] on the adjacent flood plains and was available as sulfide solution within the channel sands. Perhaps sulfate-reducing bacteria in the channel waters, fed on organic nutrients supplied from the flood plains, assisted in reducing the ferric iron. (2) The ferrous iron sulfide was dissolved in solution in carbonate water entrapped in the channel sand, which was highly porous before cementation, and expelled during compaction of the sand. (3) Ferrous iron from the expelled water was precipitated as ferric iron in a thin outer zone of the redstone beds. The relatively low permeability of the clayey, bright-red flood-plain sediments, and the contrast between oxidizing conditions on the flood plains and reducing environments in the channel waters, would facilitate the precipitation of ferric iron. Calcite and K feldspar cements in the arkoses were probably precipitated during expulsion of water by compaction. The restricted position of the bleached white areas on the borders of red arkose beds may be controlled by seepage of hydrogen sulfide from decaying plants on flood plains on either side of the stream channels.

[3] A lateritic regolith developed extensively on Precambrian, Mississippian, and older Paleozoic rocks beneath the Fountain formation suggests that the climate during Fountain deposition was warm and humid (Wahlstrom, 1948; Hubert, 1958, Ph.D. Thesis, Pa. State Univ.). Furthermore, the formation of the largely hematite-kaolinite clayey matrix of the Fountain sediments in soils on the ancestral Front Range highland required high temperature and high rainfall combined with effective leaching (Keller, 1956). Krynine (1949; 1950) reviewed thoroughly the problem of the origin of red sediments and demonstrated that, rather than implying aridity, most thick sequences of red arkoses form under wet, hot climates, frequently with a seasonal distribution of rainfall. It is well known that, except for peats, grassland soils under humid conditions yield the maximum organic content of soils (Robinson, 1951). The iron concretions, plant stems, and fossil root systems which occur in a few places in the redstone beds in the Fountain formation suggest that plant life was abundant on the flood plains. Dense plant growth probably existed on flood plains adjacent to stream channels. Subsequent rapid oxidation and destruction of the organic matter in the red sediments would occur owing to a combination of high temperature, free aeration, and sufficient moisture for microbiologic processes to operate.

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References Cited

Keller, W. D., 1929, Experimental work on red b... bleaching: Am. Jour. Sci., 5th ser., vol. 1... p. 65–70

—— 1956, Clay minerals as influenced by enviro... ments of their formations: Am. Assoc. Petro... Geol. Bull., vol. 40, p. 2689–2710

Krynine, P. D., 1949, The origin of red beds: Ne... York Acad. Sci. Trans., vol. 11, p. 60–67

—— 1950, Petrology, stratigraphy, and origin ... the Triassic sedimentary rocks of Connecticu... Conn. Geol. Survey Bull. 73, 247 p.

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